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Air Ministry
Meteorological Research Committee
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| 807<br>AD 033156 | The Free Atmosphere in the Vicinity of Fronts - Analysis of Observations by the Meteorological Research Flight (1950 - 52). | J.S. Sawyer |
| 808 | Some Examples of the Energy Spectrum of Turbulence in the Atmosphere Near the Ground. | G.D. Robinson |
| 809<br>AD 044245 | The Eddy Diffusion of Momentum, Water Vapour and Heat Near the Ground. | N.E. Rider |
| 810<br>AD 033155 | The Use of Aircraft to Measure Wind Shear by Observation of Vertical Smoke Trails. | I.J.W. Pothecary<br>R.J. Murgatroyd |
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| 820 | Five Flights Through a Thunderstorm Belt. | R.F. Jones |
| 821<br>AD 037924 | Note on the Structure of the High Altitude Strong Wind Belt in the Middle East in Winter. | J.K. Bannon |

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| 837<br>AD 063909 | The Annual and Geographical Distribution of Cold Pools. | E.J. Sumner |
| 853 | Index of M.R.P.'s Approved for General Distribution. | |

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## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

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M.R.P. No. 803 by Miss Sylvia Daniels
Mr. D.H. Johnson
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1st May, 1953.

AD 37928

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# AIR MINISTRY
# METEOROLOGICAL RESEARCH COMMITTEE

## Rainfall in relation to the jet stream

by

D.H. Johnson and Sylvia Daniels (Meteorological Office)

### 1. Introduction

There is reasonable evidence for the existence of a broad association between some types of rainfall and the jet stream. Chaudhury (1) and Yeh (2) have shown that the rainfall maxima in the mean winter precipitation patterns over India and China lie beneath the Asian low-latitude, quasi-stationary jet stream. The Himalayas complicate the Indian picture but the Chinese maximum is clearly elongated in the direction of the upper flow. Starrett (3), by comparing the locations of North American jet streams with 24 hour precipitation amounts, has shown that over America too, precipitation is greater along the jet stream than beneath the relatively slow-moving air to the north or south.

In some of the recent literature however, it is suggested that a more intimate relation between rainfall and the jet stream may exist, which depends on the preferred vertical circulation pattern within the flow. It is the purpose of the present paper to examine this suggestion with the aid of rainfall data from stations in the British Isles.

### 2. The extraction of the data

The jet streams investigated were chosen with the aid of 300 mb contour charts prepared as a six-hourly routine in the Forecast Division of the Central Forecasting Office, Dunstable, over the years 1949-51. Occasions were selected when one or more of four stations, Lerwick (60°N, 01°W), Tiree (56°N, 07°W), Valentia (52°N, 10°W), and Scilly (50°N, 06°W), were situated beneath a jet stream. Figure 1 shows the locations of each station.

The jet streams were required to be well-defined features of the upper flow with single axes: weak, complex or local wind maxima were excluded. The jet streams were each divided into three regions for the purpose of the analysis: the entrance or confluence, the centre, and the exit or difluence (Figure 2). Occasions when the flow at the axis of the jet (and also at the Station in question in the case of the central region) had appreciable curvature were excluded, i.e. portions of the jet stream which flowed around the crest of a trough or ridge were not included. No special restriction was placed on the orientation of the jet axis but since advantage was taken of every suitable occasion, there must be a strong bias towards westerly flow.

The location, relative to a jet stream, of any of the selected stations at 0300 GMT or 1500 GMT was obtained by inspection of the 300 mb contour charts and the distance of the station from the jet stream axis was measured. The position of the jet axis was estimated with the aid of continuity between the six-hourly charts. The amount of rainfall attributed to each occasion refers to the twelve-hour period centred about the time of the relevant contour chart. It was obtained from the data published in the Daily Weather Report.

### 3. Bias due to local effects

The quantity of rain precipitated at any station depends not only on dynamically induced vertical motions but also on such factors as topography, availability of moisture etc., which may be purely local characteristics Although the four control stations which were selected for this study were in

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some respects similarly situated, the mean annual rainfall varies appreciably from one to the other (see Figure 1). This variation is due, partly to local effects and partly to the locations of the stations relative to certain features of the mean atmospheric flow patterns. Local effects, however, almost certainly predominate: the distance between Scilly and Valentia is negligible on the scale of the mean circulation and yet these stations have the least and the greatest average rainfalls. It is necessary to be sure that the statistics relating to either side of the jet axis are not biased towards relatively wet or dry stations.

A convenient index of the bias due to local effects contained in the data summarised below is given by the mean of the average annual rainfalls at the four stations, weighted according to the number of occasions referring to each station. Table I contains values of the index calculated for the three sectors of the jet stream.

Table I. Indices of the bias due to local effects for each station of the jet stream.

| Sector | Left Entrance | Right Entrance | Left Centre | Right Centre | Left Exit | Right Exit |
|---|---|---|---|---|---|---|
| Index in cm | 10.9 | 11.5 | 10.5 | 10.9 | 10.3 | 10.6 |

From Table I we infer that if, for example, the rainfall distribution were in fact symmetrical about the jet axis, then we would expect the present data to show a bias of about five per cent towards the south side.

## 4. Analysis of the data

Figure 3 contains diagrams which show the average rainfall distributions relative to the jet stream axis for cases of confluence, uniform flow and difluence. The distance of the station from the jet axis was measured for each occasion to the nearest ten mi, but the position of the axis itself was certainly not known to that accuracy and so the data have been combined to give means over 100 mi steps.

Figure 4 is a composite frequency diagram giving the percentage frequency of occurrence of no precipitation, small amounts of precipitation, moderate amounts and large amounts, the terms "small", "moderate" and "large" being defined according to an arbitrary convention (small amount ≡ trace to 2.9 mm per 12 hr, moderate amount ≡ 3 to 9.9 mm per 12 hr, large amount ≡ more than 10 mm per 12 hr).

From the data summarised in Figure 3 the following features of interest emerge:

(i) In the entrance, twice as much rain occurred to the right of the axis as to the left.

(ii) In the central region, the amounts to the right and left of the axis were practically equal but, allowing for local bias, there was very slightly more to the left of the axis.

(iii) In the exit, there was twice as much rain to the left as to the right of the axis.

The frequency distributions of Figure 4 reflect the principal features of Figure 3.

## 5. Contributory causes of the rainfall pattern

The data summarised above show the average variations of rainfall from one sector to the other within the jet stream. This rainfall pattern is the consequence, not only of the mean vertical circulations but also of the average distribution relative to the jet stream of other factors, such as the degree of convective instability and the quantity of water available for precipitation.

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To obtain some idea of the effects of convective instability in the above analysis, the data for the central and exit regions (there being insufficient data for the entrance) was further classified according to the character of the precipitation. The nature of the precipitation was determined from the descriptions of the weather, in terms of Beaufort letters for the appropriate twelve-hour period, contained in the Daily Weather Report. Three classes were formed (a) when showers only were reported, (b) when both rain and showers were reported for the same twelve-hour interval, (c) when rain alone was reported. The distribution of rainfall for each class is shown in Figure 5.

Since the number of cases comprising each class in the exit region is small, the distributions have been smoothed by taking mean values for overlapping three hundred mile intervals. Figure 5 shows that rain type precipitation increased very markedly towards the left of the difluence (there is a five per cent bias towards the south due to local effects). So the elimination of precipitation of a showery type does not affect the rainfall distribution in the exit region.

The preponderance of rainfall to the right of the confluence and to the left of the difluence of the flow is indicative of mean vertical circulations within the current in the sense required by the flow accelerations and the associated lower tropospheric developments required according to Sutcliffe's theory (4). More explicitly: the acceleration of the flow at the confluence sets up an ageostrophic motion, due to an unbalanced component of the pressure force, from right to left across the jet axis in the upper troposphere, with convergence to the left and divergence to the right of the axis. A compensating return flow, with divergence to the left and convergence to the right, occurs in the lower troposphere. To complete the cross-axis circulation model, upward vertical motion is required to the right of the axis and downward vertical motion to the left. This type of vertical circulation, occurring in the direction of natural convective overturning has been referred to as a "direct" circulation. A cross-axis circulation in the opposite ("indirect") sense should occur at the difluence. These effects have been discussed in detail by several writers, perhaps most recently by Murray and Daniels (5), who have demonstrated statistically the existence of the appropriate cross-axis flows at jet stream levels. The present work may be taken as independent evidence for the associated vertical motions. We also note that showers to the left of the axis in the exit gave more precipitation than those which occurred to the left of the axis in the central region, which suggests that convective precipitation at the left exit of the jet may have been intensified by large scale convergence.

From Figure 5 we see that in the central region, rainfall amounts due to both shower and mixed types of precipitation increase towards the left of the axis, i.e. in the direction in which convectional instability increases. The class for which dynamical effects are likely to be most evident is that which consists only of cases where precipitation was reported as rain. For these cases rainfall amounts are, if anything, greater to the right of the axis. However, there is a seven per cent bias towards the right due to local effects and some allowance must be made for the cross-flow gradient of humidity mixing-ratio, which for the same vertical lifting would result in the precipitation of a greater quantity of moisture to the right than to the left of the current. We consider that the data is inadequate to decide whether there is any preferred vertical circulation within the central region of the jet stream.

Most of the moisture available for precipitation is contained at levels below 500 mb, so vertical motions in the lower troposphere are of more direct importance for precipitation than vertical motions which occur in the upper troposphere. Jet streams are essentially features of the upper troposphere and lowest stratosphere and quasi-uniform jets can be associated with a diversity of lower tropospheric features; both dynamical theory and synoptic experience teach that such jet streams act as steering currents for anticyclonic as well as cyclonic activity. Evidently uniform upper currents can exist in combination with a variety of lower tropospheric vertical circulation patterns: the lack of any clear cut association of rainfall with one side or the other of the central region of the jet stream does not necessarily imply the presence of a preferred vertical circulation pattern at jet stream levels.

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## 6. Conclusions

The mean rainfall patterns at the entrance and exit of the jet stream imply mean cross-axis vertical circulations with ascent beneath the right entrance and left exit, and descent beneath the left entrance and right exit. These circulations agree with accepted dynamical theory. No definite conclusions with regard to preferred vertical circulations within the central region of the jet stream can be made from arguments based purely on the mean precipitation patterns found for uniform currents.

## 7. References

(1) Chaudhury, A.M. 1950 Tellus, Vol. 2, p.56.

(2) Yeh, T.C. 1950 Tellus, Vol. 2, p.173.

(3) Starrett, L.G. 1949 J. Meteor., Vol. 6, p.347.

(4) Sutcliffe, R.C. 1947 Quart. J., R. met. Soc., Vol. 73, p.370.

(5) Murray, R. and Daniels, S.M. 1951 M.R.P. 690.

## LOCATION OF THE OBSERVING STATIONS

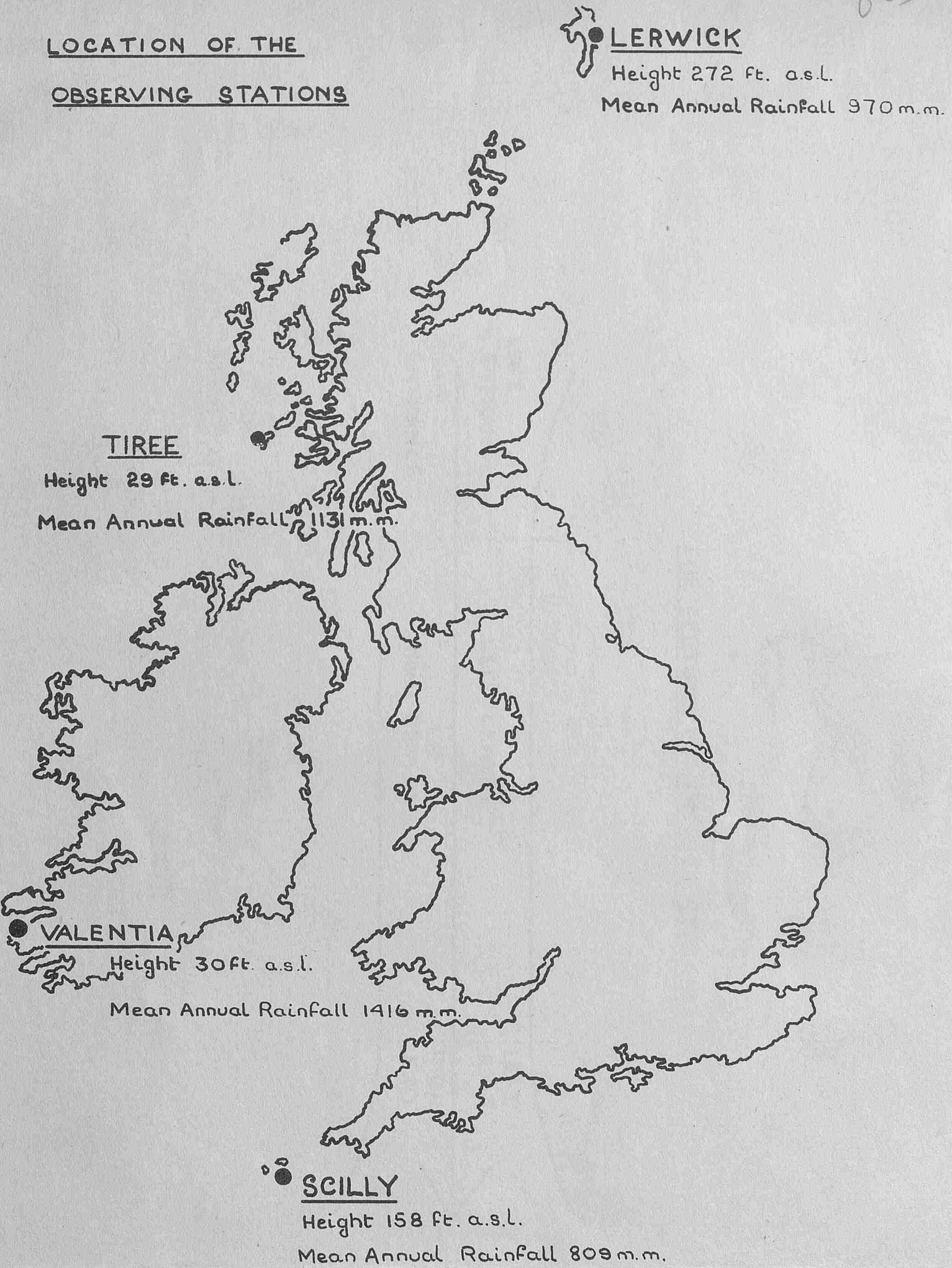
LERWICK
Height 272 ft. a.s.l.
Mean Annual Rainfall 970 m.m.
TIREE
Height 29 ft. a.s.l.
Mean Annual Rainfall 1131 m.m.
VALENTIA
Height 30 ft. a.s.l.
Mean Annual Rainfall 1416 m.m.
SCILLY
Height 158 ft. a.s.l.
Mean Annual Rainfall 809 m.m.

FIG. 1.

## THE SECTORS OF THE JET STREAM

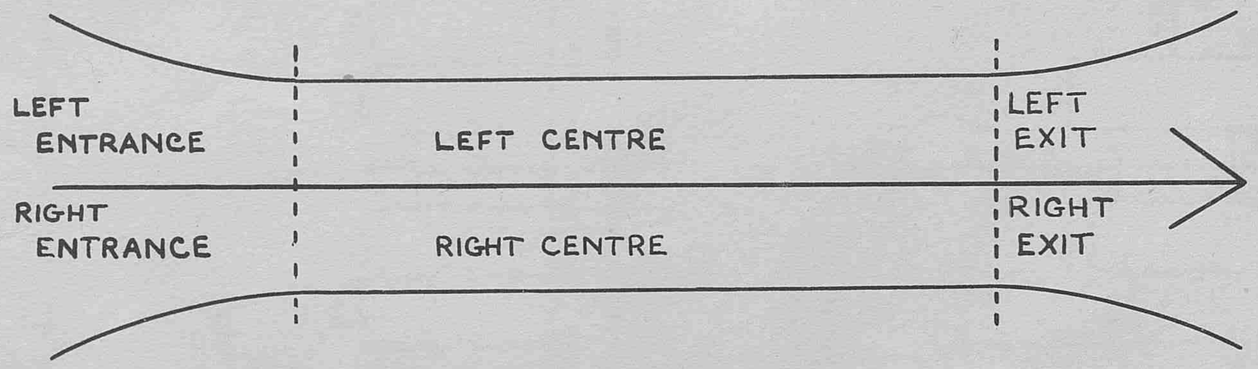
LEFT ENTRANCE
RIGHT ENTRANCE
LEFT CENTRE
RIGHT CENTRE
LEFT EXIT
RIGHT EXIT

FIG 2

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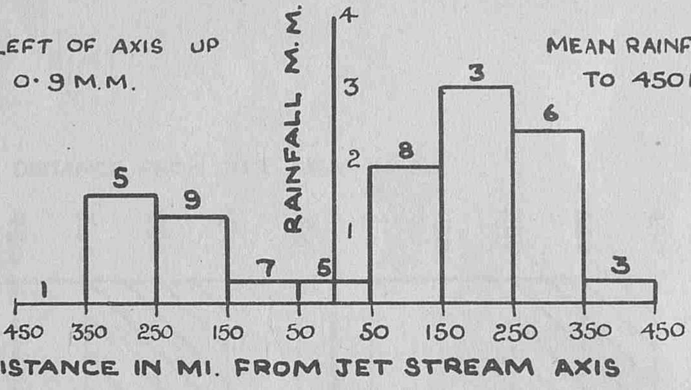
# DISTRIBUTION OF RAINFALL BENEATH THE JET STREAM.

## ENTRANCE

MEAN RAINFALL LEFT OF AXIS UP TO 450 MI = 0·9 M.M.

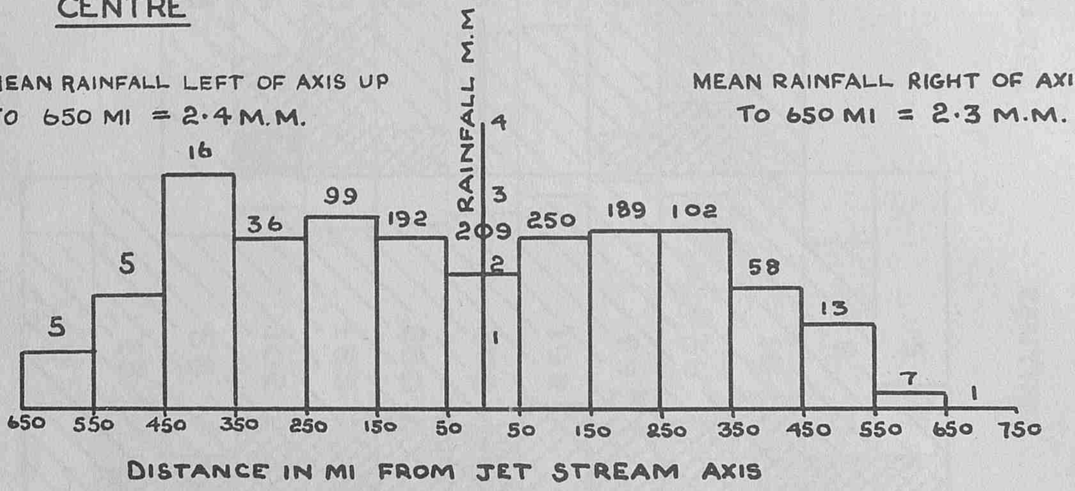
MEAN RAINFALL RIGHT OF AXIS UP TO 450 MI = 1·8 M.M.



## CENTRE

MEAN RAINFALL LEFT OF AXIS UP TO 650 MI = 2·4 M.M.

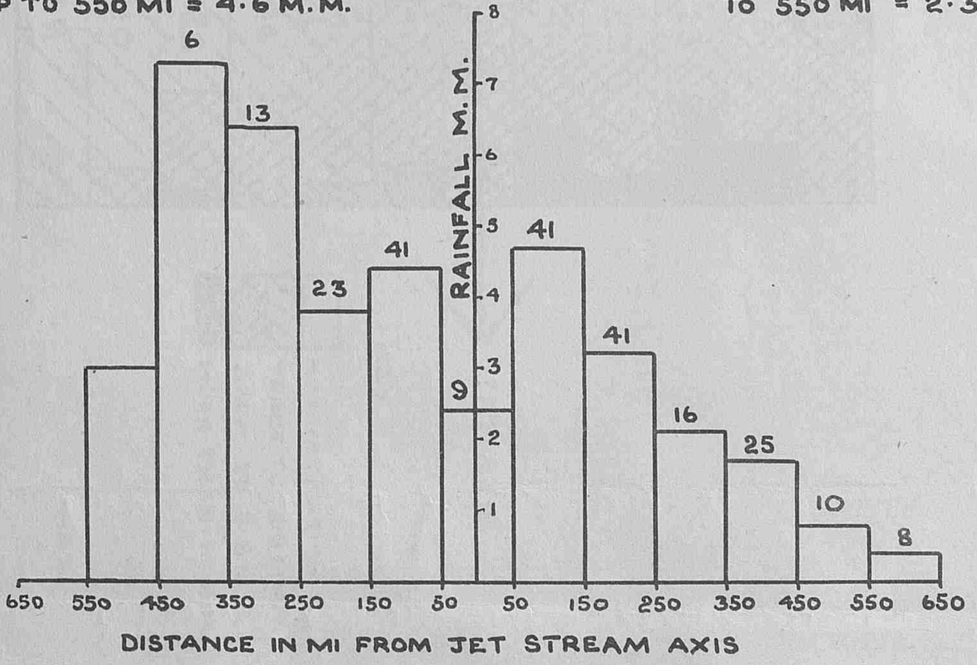
MEAN RAINFALL RIGHT OF AXIS UP TO 650 MI = 2·3 M.M.



## EXIT

MEAN RAINFALL LEFT OF AXIS UP TO 550 MI = 4·6 M.M.

MEAN RAINFALL RIGHT OF AXIS UP TO 550 MI = 2·3 MM



FIGURES ABOVE THE DIAGRAMS GIVE THE NUMBER OF CASES FOR EACH INTERVAL

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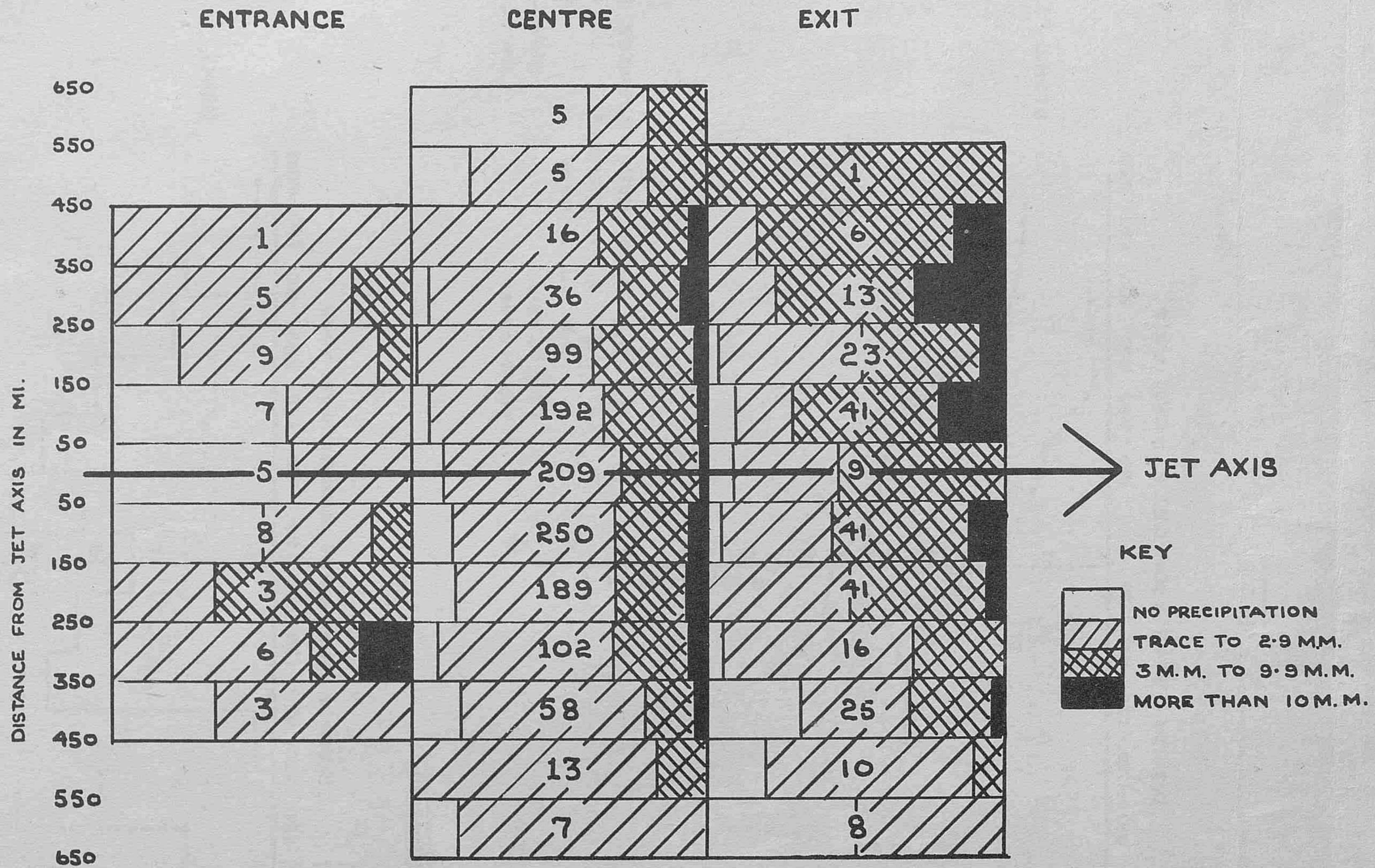
FIG 3

## COMPOSITE FREQUENCY DIAGRAM OF PRECIPITATION IN THE JET STREAM



THE SHADED AREA IS PROPORTIONAL TO THE PERCENTAGE FREQUENCY
THE ENCLOSED FIGURES GIVE THE NUMBER OF CASES IN EACH DIVISION

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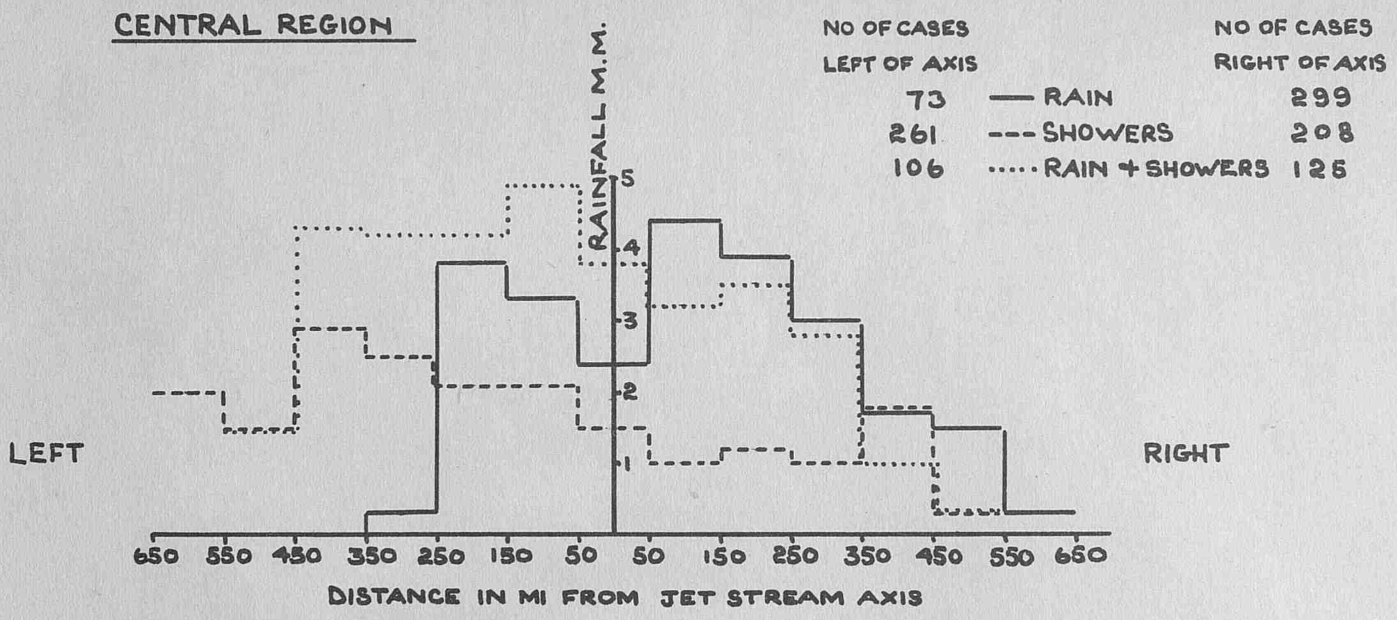
FIG 4.

VARIATION OF THE MEAN RAINFALL DISTRIBUTION WITH PRECIPITATION TYPE.

CENTRAL REGION



EXIT REGION

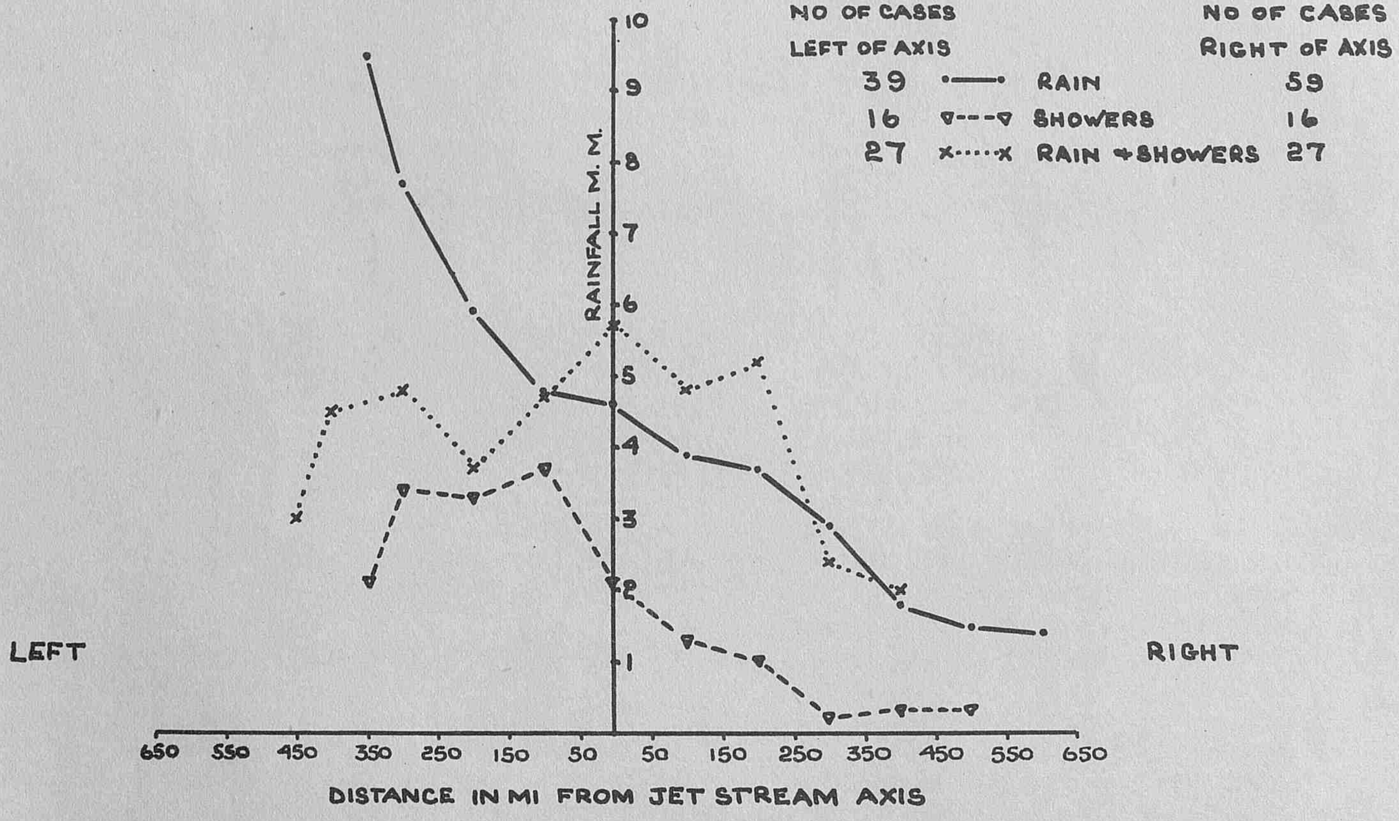


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FIG 5

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M.R.P. 804
S.C.I/78
29th April, 1953

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METEOROLOGICAL RESEARCH COMMITTEE

# Hot-Wire Anemometers for Use on Aircraft

by

R.J. Murgatroyd
D.G. James
(Meteorological Office)

## 1. Summary.

A description is given of a hot-wire anemometer designed to measure the gustiness of the airstream through which an aircraft is flying. Measurements of gust amplitudes down to about 2 feet per second and frequencies up to 50 cycles per second are obtained and recorded by photographing a cathode ray tube display.

By means of a modified instrument gust components perpendicular to the direction of motion of the aircraft can be obtained and this has been used to measure vertical gusts in flight.

Design and circuit details are given and typical flight records are shown.

## 2. Introduction.

Any attempt to examine the properties of atmospheric turbulence either at the earth's surface or in the free air must necessitate the measurement of both wind and temperature fluctuations in the horizontal and vertical. To obtain as true a picture as is possible, the instruments must have a very rapid response compared with the speed of the fluctuations which are to be investigated.

Grant (1950) Ref.1 designed an ultra-rapid thermometer, for use in aircraft, consisting of several 50 s.w.g. platinum wires each some 5 cm. long, and measured their change of resistance accurately to give the fluctuations of the temperature of the ambient air. The lag of this instrument is about 4 milliseconds and it is therefore capable of measuring fluctuations up to 250 cycles per second.

The circuit arrangement of the ultra-rapid thermometer includes a low frequency oscillator with its output connected to a balanced Wheatstone bridge having as one of its arms the temperature sensitive resistance. The output from the bridge after amplification and rectification is applied to a cathode ray tube display which can be photographed. It appears feasible to replace this bridge by another bridge with a hot-wire anemometer element as one of its arms. Once this is achieved it becomes possible to measure gustiness from an aircraft and by duplication of the apparatus to correlate it with temperature fluctuations in the free air.

## 3. Aims.

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## 3. Aims.

Hot-wire anemometers are used extensively in turbulence measurements on the surface and in wind tunnels. The literature on this aspect of their use is considerable and recent meteorological applications are described in Refs.2 Swinbank (1948) and 3 Lander & Robinson (1952). Their use on aircraft introduces difficulties with regard to their mechanical strength and their ability to stand vibration. In addition the implications of their location on a power driven moving body have to be considered.

/A

A large aircraft possesses considerable inertia and its response time to a sudden gust may be as much as 5 sec. In an aircraft travelling at a speed of 120 knots this represents a movement of 1000 feet or so. Hence for all turbulence phenomena with wavelength say below 100 or 200 feet the aircraft may be considered as a steady platform. It has been found possible to design a hot-wire anemometer strong enough to work on an aircraft and yet able to record frequencies up to 20 or possibly 50 cycles per second i.e. of measuring phenomena with wavelengths down to about 10 feet. If the aircraft is flown at constant power and "hands off" the height and airspeed meters will give an indication of the long wave fluctuations while the hot-wire anemometer will record faithfully any fluctuations in the air-stream relative to the aircraft with wavelength between say 10 and 200 feet.

A simple heated wire will respond to both the temperature of the air and to the wind components in planes perpendicular to its length. Devices to eliminate the effect of temperature fluctuations and also to measure wind components in given directions have been described elsewhere Ref.4 Swinbank (1951) and Ref.5 Simmons and Bailey (1927) for surface instruments. Once the simple hot-wire anemometer has been made to operate on an aircraft the application of these modifications to the aircraft instrument is comparatively straightforward and two examples of this are described below. The second of these which is capable of measuring gust velocities in the upwards and downwards directions independently of the speed of the aircraft is of particular importance in investigations of bumpy conditions of flight.

## 4. Design and construction.

### (i) The temperature compensated hot-wire anemometer.

It has been demonstrated by Grant (1950) Ref.1 that 50 s.w.g. platinum wire can be exposed from an aircraft travelling at speeds up to at least 150 knots without breakage provided that precipitation and icing conditions are avoided. Using this wire as the sensitive element his ultra-rapid thermometer has a lag of about 4 milliseconds. When this wire is heated considerably for use as a hot-wire anemometer however it loses tensile strength and after tests in a wind tunnel it was decided to use 48 s.w.g. platinum wire for this purpose. This increased the diameter of the wire two fold and hence the lag of the instrument four fold or somewhat more, but this is still acceptable. An alternative proposal of using a wire of higher tensile strength e.g. nickel was abandoned after tests had shown that it quickly changed its calibration with use.

It is shown in Ref.6 (Schuh,H.) that as the temperature of a hot wire is raised its sensitivity to wind fluctuations increases considerably compared with its sensitivity to temperature fluctuations. The wind tunnel tests made with the 48 s.w.g. platinum wire over a temperature range -20°F to + 60°F and a speed range of 100 to 140 knots gave the following equivalent sensitivities.

| Wire Current. | Temperature change to produce same voltage change as one knot change in wind speed. |
|---|---|
| 1 amp | 2.0°F |
| 1.1 amp | 4.4°F |

Hence it is a considerable advantage to operate the hot wire at as high a temperature as possible. This particular instrument is arranged so that it becomes a dull red (circa 550°C) at about 100 knots airspeed and + 60°F ambient temperature. The aircraft speed when it is in use is always between 115 and 120 knots at present. Theoretical calculations suggest that a change of ambient temperature of 1°F will change the wire temperature by about 1°C. It is thought therefore that the wire temperature in use is usually between 400 and 450°C. It was observed in the wind tunnel that any increase in current resulting in a change from dull red to bright red heat was quickly followed by a breakage of the wire.

Further improvement in the sensitivity to wind fluctuations compared to the sensitivity to temperature fluctuations can be obtained by using a bridge circuit of the type shown in Fig. 1. By arranging that the current is large through the platinum hot-wire anemometer and small through the platinum thermometer element, both elements respond to temperature fluctuations whilst one only responds to wind fluctuations. The bridge output is therefore dependent on wind fluctuations only. As a small design point, allowance can be made for the departure from linearity of the variation of the resistance of platinum wire with temperature, (which becomes significant for the hot wire), by inserting a small resistance which is not sensitive to temperature in series with the compensating element. The actual values of the components used are given in the figure. Balance is obtained by means of the potentiometer in the high impedance side of the bridge (see Fig. 5). Other features of this bridge were introduced to fit it into the oscillator and amplifier circuits of the ultra-rapid thermometer and these are described in a subsequent paragraph. The power supply for the hot wire is a separate 24 volt accumulator.

This instrument was mounted on the aircraft and made retractable into the aircraft skin by adapting the head of the ultra-rapid thermometer. This became possible as the circuit modifications described below greatly increased the sensitivity of the apparatus and made it possible to dispense with some of the ultra-rapid thermometer wires. A photograph of the new combined hot-wire anemometer - ultra-rapid thermometer head is shown in Fig. 2. The two hot wires are each about 2 cm. long and connected in series. They are easily soft-soldered into position care being taken to make them as tight as possible between their supports. To save space the temperature compensating wires are made of 50 s.w.g. platinum wire similar to those used in the ultra-rapid thermometer.

(ii) The directional hot-wire anemometer.

Directional hot-wire anemometers used on the surface (Refs. 4, 5 & 6) usually consist of two hot wires inclined at equal angles above and below the horizontal and connected in opposed arms of a Wheatstone bridge. For horizontal wind components the bridge remains in balance while vertical wind components cool one wire more than the other and hence unbalance the bridge. No temperature compensation is necessary as temperature fluctuations will be in phase in the opposed arms of the bridge.

The same result may be obtained by mounting small hot-wire elements within two tubes inclined at similar angles above and below the horizontal. An arrangement of this type appeared preferable for aircraft work and some preliminary wind tunnel tests were carried out along these lines. It was found that the larger the angles of inclination to the horizontal the greater the sensitivity of the arrangement but the greater the disturbance to the airflow produced by the tubes. An angle of 45° is probably the best compromise.

The final design of the head adapted for aircraft use is illustrated in Fig. 3.

This figure is not exact in that the airfoil depicted is not quite the same shape as the section employed in the head, and has been drawn thus to permit easier understanding of its action. The drawing is also not to scale, the length of the actual head being about 8 inches, its vertical dimension being 3.5 inches, whilst its depth is 2.5 inches. The supports for mounting on the aircraft are also not shown, to avoid confusion, but can easily be seen in Fig. 4.

The left hand diagram in Fig. 3 is a longitudinal section through an airfoil in which the hot wires are mounted. The hot wires are perpendicular to the plane of the paper and the air flow between the inner and outer shells of the airfoil is shown. If the flow is horizontal it will affect the upper and lower hot wires equally. If it has, say, an upward component the flow through the upper channel will be greater than that through the lower channel.

/This

This head is mounted about 5 feet ahead of the nose of the aircraft. In precipitation or icing conditions the wires can be protected by rotating the shaft. This moves the inner nose forward to act as a valve and close up the entrance to both channels. A photograph (Fig. 4) shows the head mounted on the aircraft. Means are also provided to rotate, relative to the aircraft, the arm supporting the head for calibration purposes and this feature is described in a subsequent paragraph. The airfoil is made of perspex except for a small paxolin frame on which the wires are mounted. This frame may easily be withdrawn for maintenance purposes.

The wires are connected in the Wheatstone bridge circuit shown in Fig. 6. There are two hot wires in series in each channel each about 3 cm. long and hence there is a resistance of about 4.5 ohms in each of the sensitive arms of the bridge. This means that the bridge presents a very low impedance to the oscillator output and this and other features presented circuit problems regarding the adaptation of the bridge to the ultra-rapid thermometer apparatus.

## 5. Modifications to the ultra-rapid thermometer circuits.

To adapt the circuitry and display system used on the ultra-rapid thermometer it is first necessary to feed the 1800 c/s signal from the oscillator into the bridge. Since the bridge (Fig. 5) has a low impedance, matching from the anode of the oscillator to the bridge must be effected by the use of a tuned transformer. In fact two transformers were used giving in all a turns ratio of some 300: 1, $T_1$, tuned with a 0.1 μ F condenser, for maximum power output of the oscillator.

Similarly, the output of the bridge is matched and tuned into the grid circuit of the first amplification stage, for maximum A.C. volts. Correctly matched, there is no need for the large amplification available from the original ultra-rapid thermometer amplifier, and consequently decoupling of the cathode circuits may be decreased to produce greater stability. This was found to be necessary since large amplitude self-oscillation in the amplifier could be produced by slight mechanical vibration, if the bridge circuit was not correctly matched to the first stage.

Two 250 μ F condensers prevent any D.C. flowing through the slide wire potentiometer and the primary winding of the output transformer, $T_2$, whilst a 2 μ F condenser isolates the secondary of transformer $T_1$. Another variable condenser, C, is used to obtain perfect balance of the bridge, since the stray capacities in the wiring introduce a small phase change in the output voltage vector.

## 6. The effect of aircraft vibration.

Aircraft vibration gives rise both to mechanical movement of the elements and also to some change in electric circuit components. With the low impedance bridges and high gain amplifiers used the latter problem is the more serious. In particular switch contacts and moving contacts on potentiometers give considerable trouble and must be located at low current points as far as possible and maintained frequently. Antivibration mountings and screened leads are essential. In order to reduce any spurious voltages the oscillator waveform was improved while electrical pick up by the various exposed leads was reduced by correct tuning of the bridge input transformer and the introduction of a narrow band pass filter in the input to the first amplifier stage.

Mechanical vibration of the simple anemometer lead is not a serious problem. It is suspended on a rod of elliptical cross-section with major axis in the line of flight. Hence any vibration of the rod will tend to take place in the plane of the hot wires and thus will have a very small effect on their cooling. In the directional head however the present supporting structure vibrates in flight and produces a noise level which is somewhat higher, although not unacceptable. It is hoped to reduce this further at a later stage by redesigning this structure.

/7.

## 7. Flight procedure.

The flight procedure for the simple hot-wire anemometer is as follows:-

(a) All the power supplies to the amplifiers and cathode ray tube assembly are switched on.

(b) The head, on which are situated the hot wires and the temperature compensating wires, is lowered from its protective fairing below the nose of the aircraft, and the rod clamped into position. While this is done the aircraft is flown at as low a speed as possible in clear air.

(c) The D.C. accumulator supply is connected to the hot wires, and the current through them brought to the desired value, for a given airspeed, by use of the rheostat.

(d) The bridge is balanced both for resistance and capacity by observing the spot on the cathode ray tube, and the desired range of sensitivity is selected.

(e) The pilot is requested to keep a constant airspeed as far as possible, and the recording cameras are switched on. It is essential to avoid precipitation and icing conditions in flight but the wires do not usually break in thin water cloud.

The directional hot-wire anemometer is operated similarly in flight. It has been arranged to use the same components as far as possible in the two bridges and by means of a multiple switch a change from one instrument to the other can be effected immediately. When this instrument is in use the pilot of the aircraft is requested to fly at constant attitude.

## 8. Calibration.

The calibration of both instruments is carried out after every flight on which they have been used. For the simple hot-wire anemometer the aircraft, flying straight and level in non-turbulent air, is made to increase speed slowly, when a given deflection of the spot on the cathode ray tube is measured directly in knots from the indicated airspeed meter on the aircraft. For the instrument described above, sensitivities of 20 knots full scale to 1 knot full scale are available.

For the calibration of the directional hot-wire anemometer use is made of the arrangement by which the rods supporting the sensitive head rotate on a pivot fixed in the aircraft's nose. By turning a screw the head can be inclined to or depressed from the incident airstream (Fig. 7). Thus by measuring the angle of inclination or depression and the airspeed, the effective vertical wind component for a given movement of the spot on the cathode ray tube can be determined.

## 9. Typical records.

Figs. 8, 9 & 10 show traces obtained from the simple hot-wire anemometer and the directional hot-wire anemometer respectively during a flight in the friction layer on 25/8/1952. A trace from the ultra-rapid thermometer instrument is also shown as well as an accelerometer trace corresponding to the directional hot-wire anemometer record. It is apparent that these instruments are capable of measuring gust speeds down to 2 ft/sec. and frequencies up to 50 cycles per second.

The records obtained from the directional hot-wire anemometer and those of the recording accelerometer show good agreement in the size and duration of vertical gusts producing bumps, as recorded by these traces. It can be seen that for a large duration gust, the aircraft will eventually follow it, and so the directional hot-wire anemometer gives us in the first instance, the sharpness of the edge of any vertical gust. Incidentally one can also obtain its shape and duration, for the aircraft acts as a differentiating circuit for gusts of wavelength greater than or equal to the linear dimension of the aircraft.

## 10. Discussion.

The two instruments described above could be used in conjunction with a third head directional in the horizontal plane of the aircraft's flight to obtain all three components of a gust vector wind. The first requirement in this work however is simultaneous recording of the outputs of these various anemometers with that of the ultra-rapid thermometer. It is probable that this can be achieved with a multi-channel recording galvanometer or alternatively the present cathode ray tube could be quadruplicated and a suitable recording cathode ray tube camera built.

Meanwhile, until simultaneous recording can be achieved, it is possible to obtain limited data on turbulence and also convective processes in the atmosphere by using the thermometer and anemometer separately in turn.

## References.

1. Grant, D.R. 1950. M.R.P. 542.
2. Swinbank, W.C. 1948. M.R.P. 429.
3. Lander, A.J. and G.D. Robinson. 1952. M.R.P. 740.
4. Swinbank, W.C. 1951. J.Met. 8, p.135.
5. Simmons, L.F.G. and A. Bailey. 1927. Phil. Mag., Series 7. 3.
6. Schuh, H. "Measurements of wind fluctuations by means of hot wires". Vӧlkenrode. R. & T. No. 947.

Fig 1. Simple Bridge for Hot-Wire Anemometer.

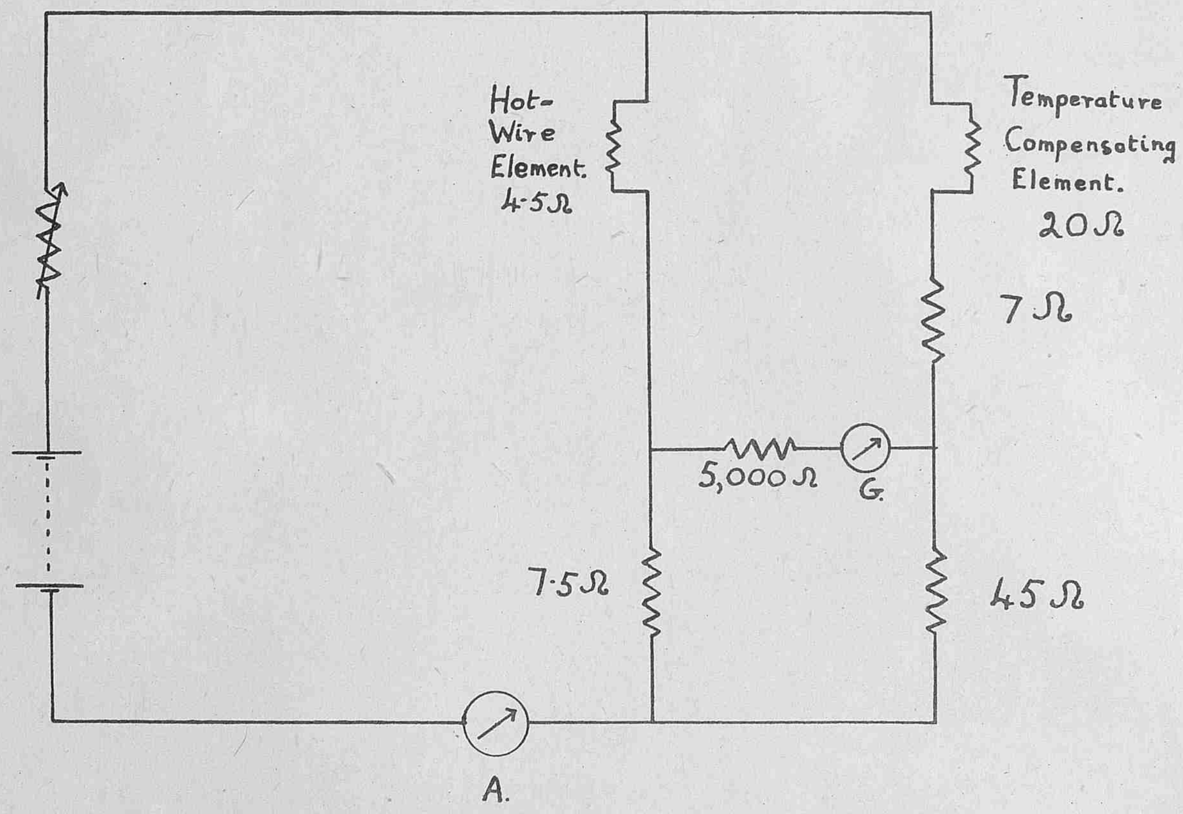

Hot-Wire Element. 4·5Ω
Temperature Compensating Element. 20Ω
7Ω
5,000Ω
G.
7·5Ω
45Ω
A.

FIG 2. SENSITIVE HEAD INCORPORATING THE HOT-WIRE ANEMOMETER AND ULTRA-RAPID THERMOMETER.

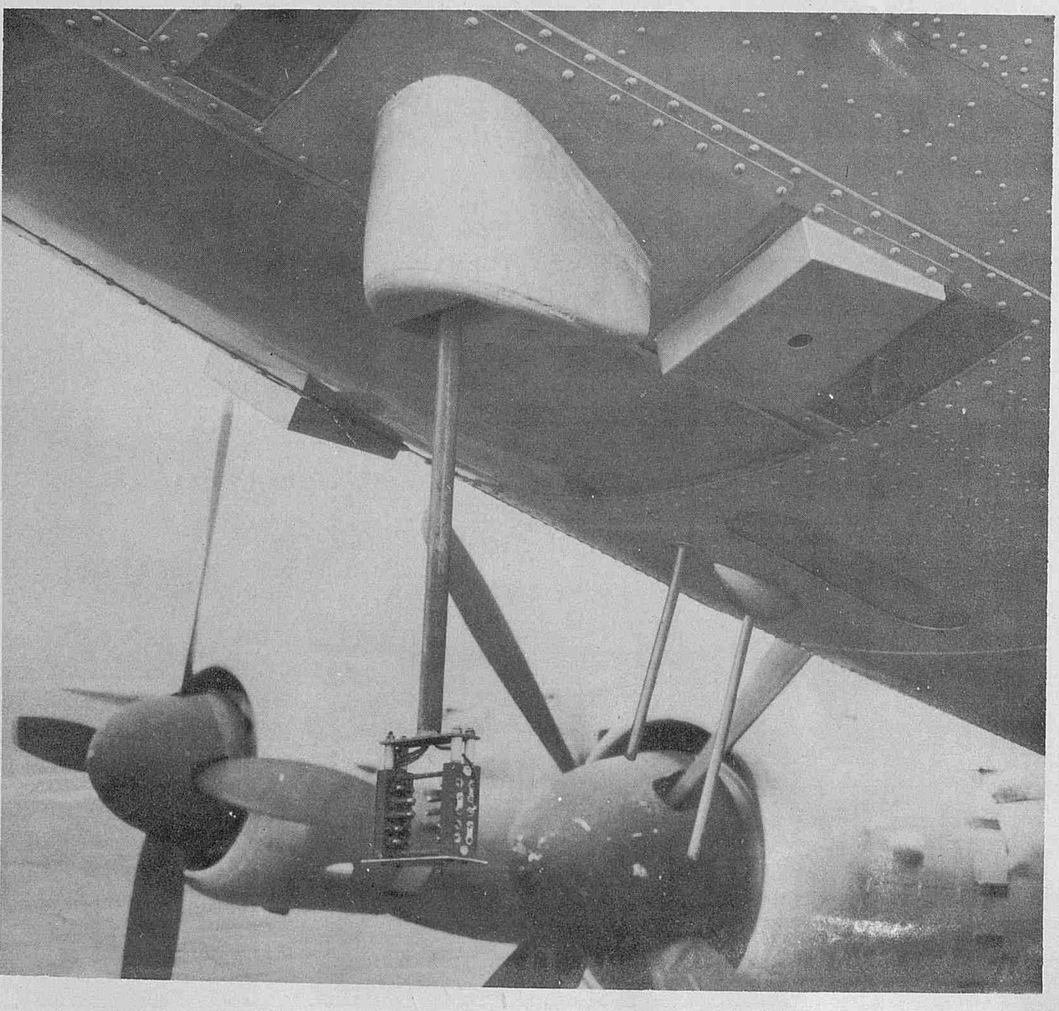
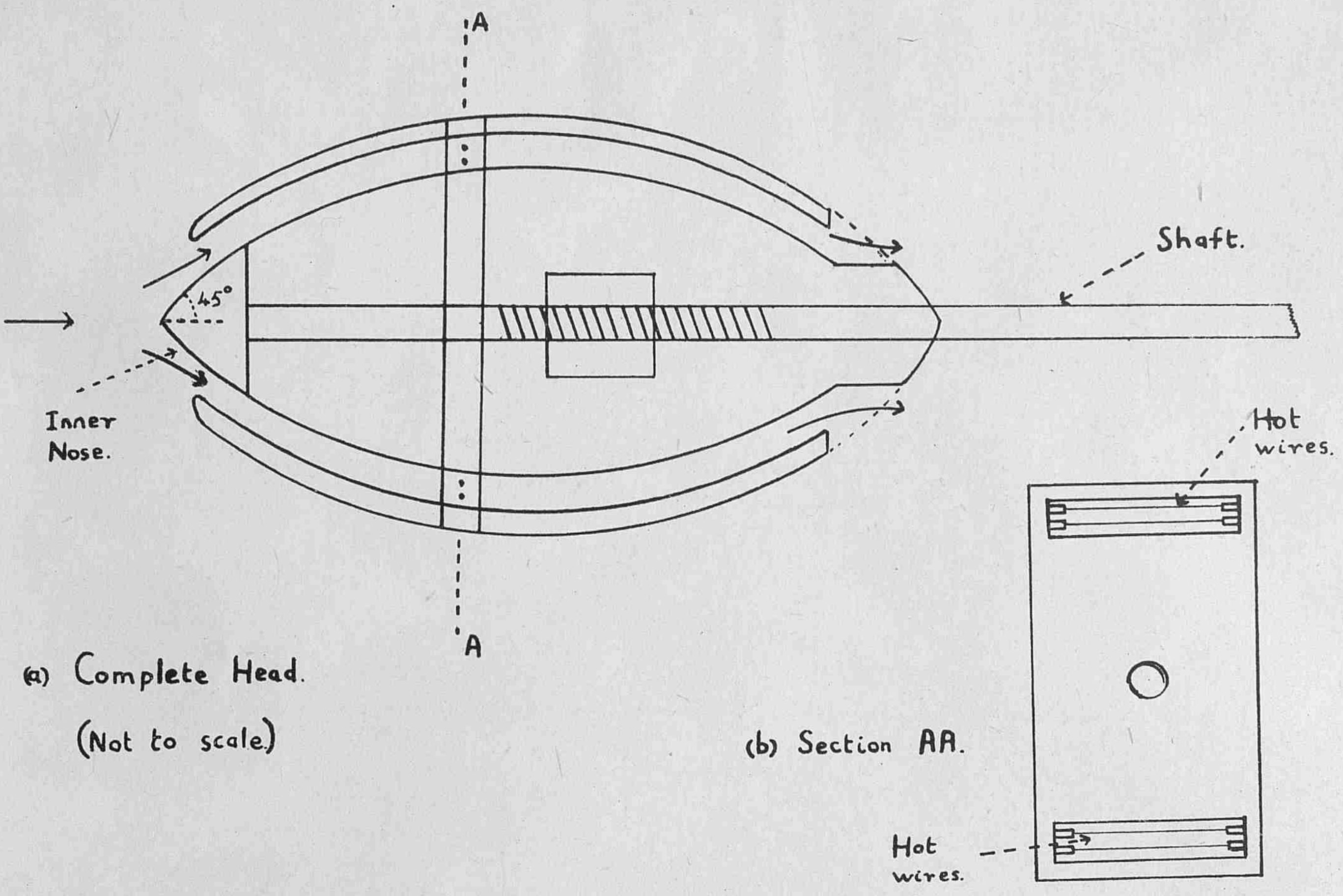


Fig 3. Directional Hot-Wire Anemometer.


A
45°
Shaft.
Inner Nose.
A
Hot wires.
Hot wires.

(a) Complete Head.

(Not to scale)

(b) Section AA.

FIG. 4. DIRECTIONAL HEAD MOUNTED ON AIRCRAFT.

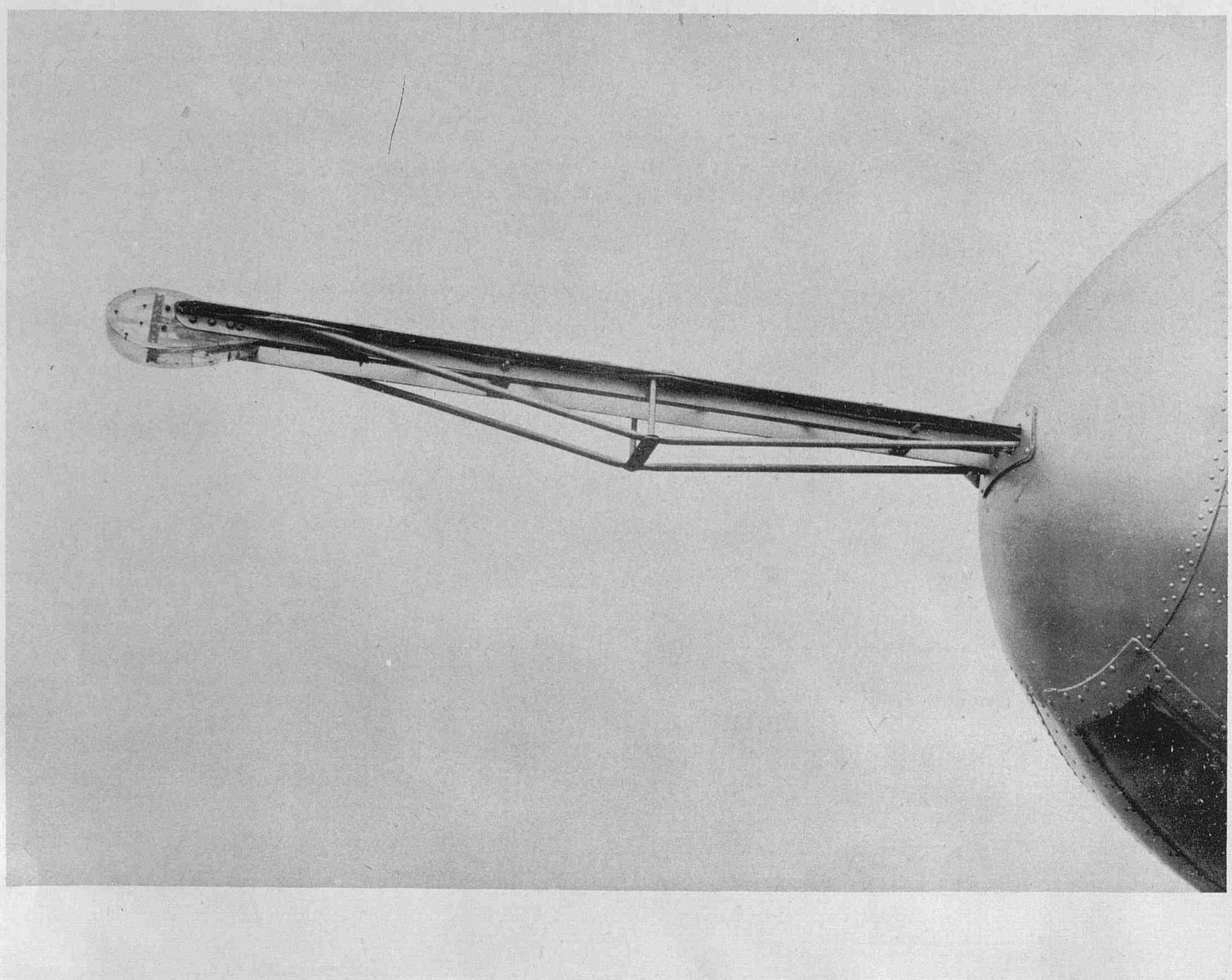


Fig 5. Bridge for Hot-Wire Anemometer.

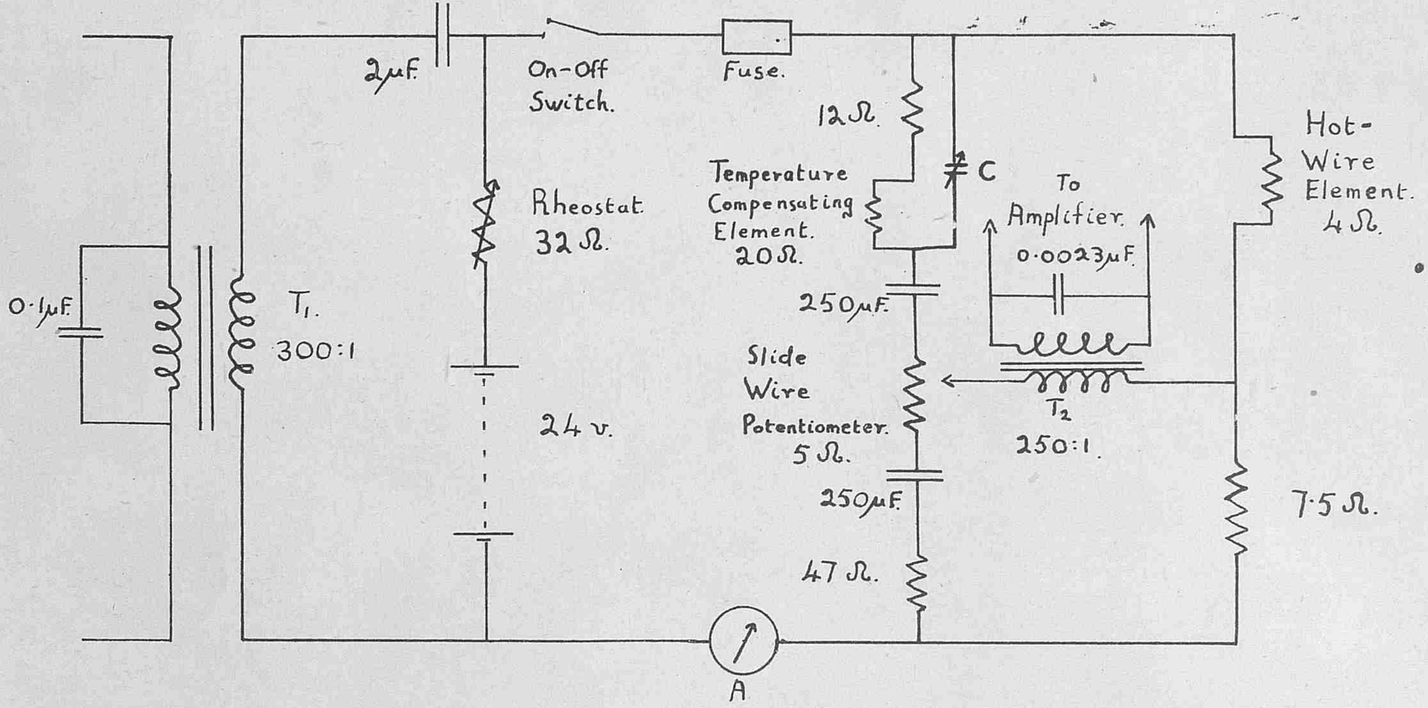

2μF.
On-Off
Switch.
Fuse.
12Ω.
Hot-
Wire
Element.
4Ω.
C
To
Amplifier.
0·0023μF.
Rheostat
32Ω.
Temperature
Compensating
Element.
20Ω.
250μF.
0·1μF.
T₁.
300:1
Slide
Wire
Potentiometer.
5Ω.
T₂
250:1.
24v.
250μF.
7·5Ω.
47Ω.
A

Fig 6. Bridge for Directional Hot-Wire Anemometer.

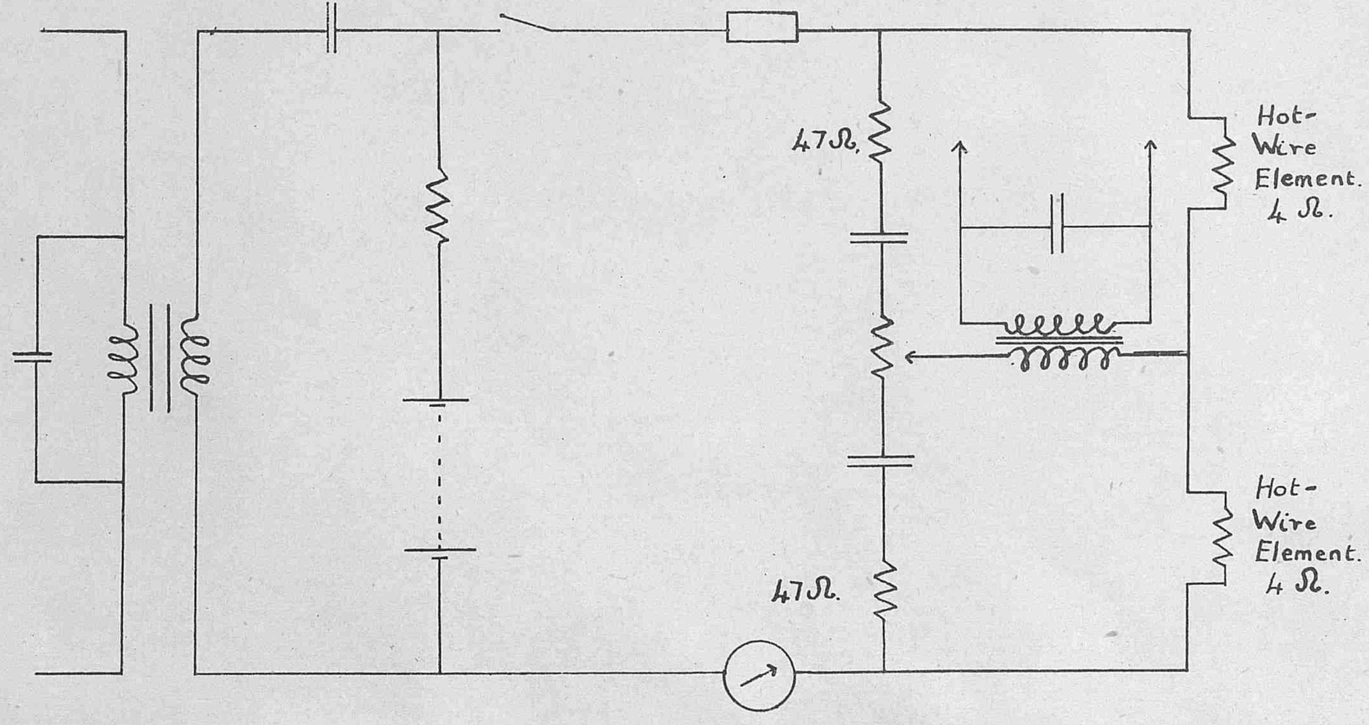
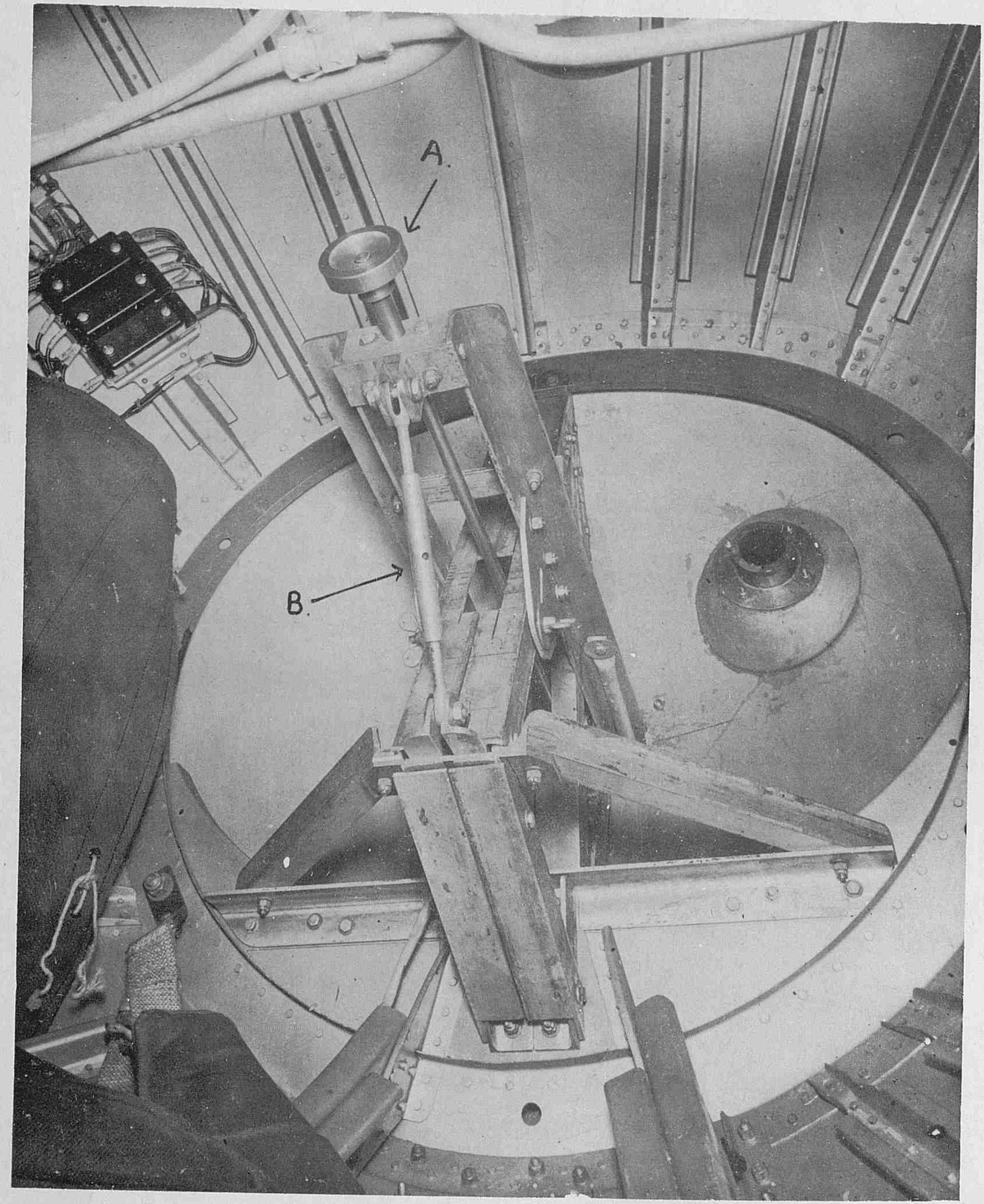

47Ω.
Hot-
Wire
Element.
4Ω.
Hot-
Wire
Element.
4Ω.
47Ω.

FIG 7. CLAMPING ARRANGEMENT INSIDE AIRCRAFT FOR DIRECTIONAL HEAD SUPPORTS.

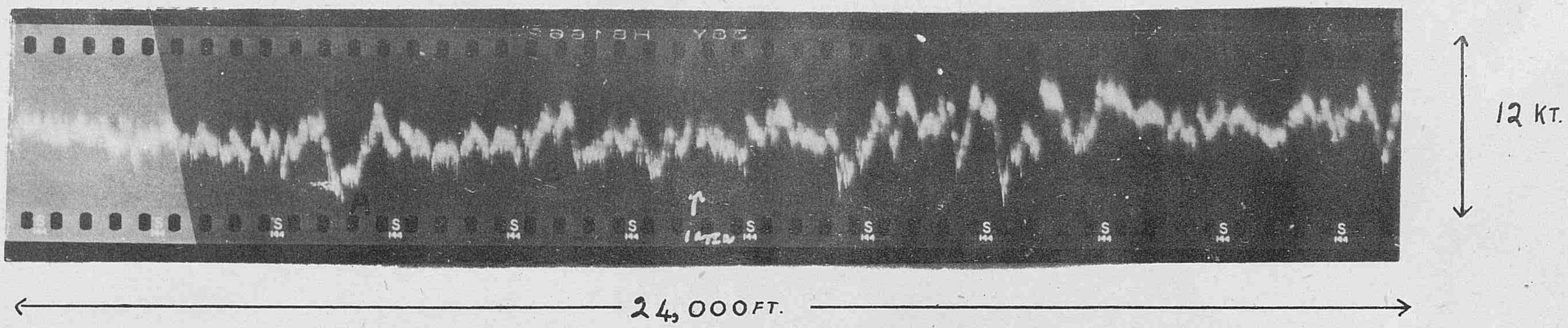


A. ROTATING ROD FOR OPENING AND CLOSING THE NOSE OF THE SENSITIVE HEAD.

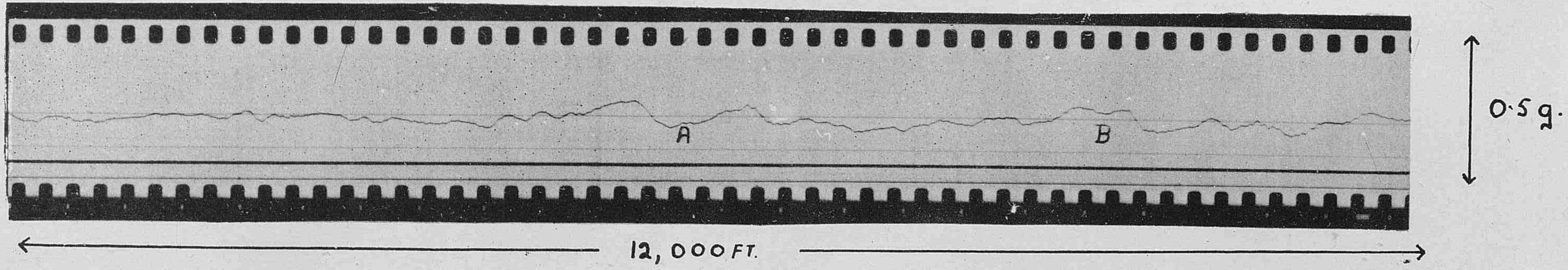
B. ADJUSTABLE SCREW FOR CALIBRATION.

FIG 8(a). RECORD OBTAINED FROM DIRECTIONAL HOT-WIRE ANEMOMETER.

25TH. AUGUST 1952 AT 1,500 FT.



(b) ACCELEROMETER RECORD CORRESPONDING TO FIRST HALF OF ABOVE TRACE.



M.R.P. 804

FIG. 9. RECORD OBTAINED FROM SIMPLE HOT-WIRE ANEMOMETER.

25TH. AUGUST 1952 AT 1,500FT.

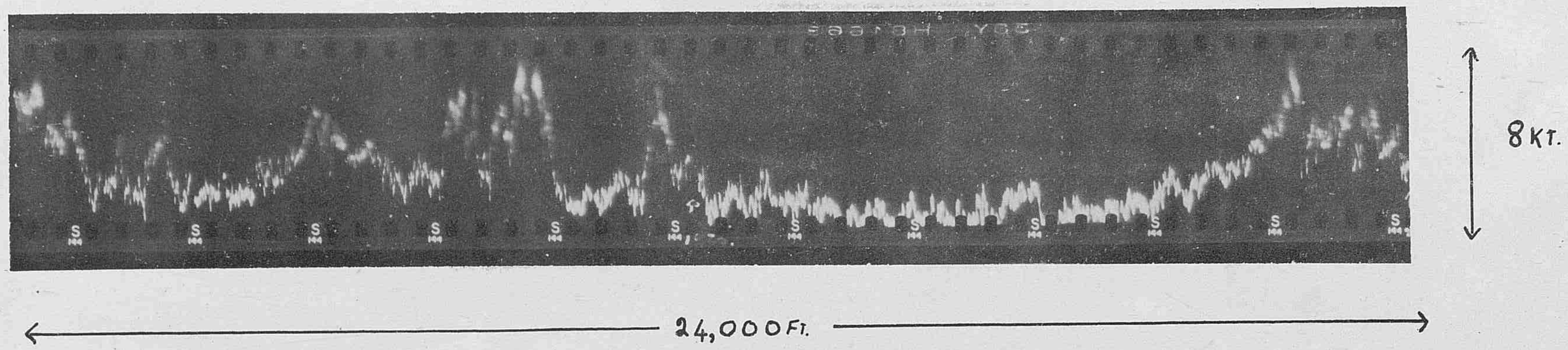


FIG 10. RECORD OBTAINED FROM ULTRA-RAPID THERMOMETER.

25TH. AUGUST 1952 AT 1,500FT.

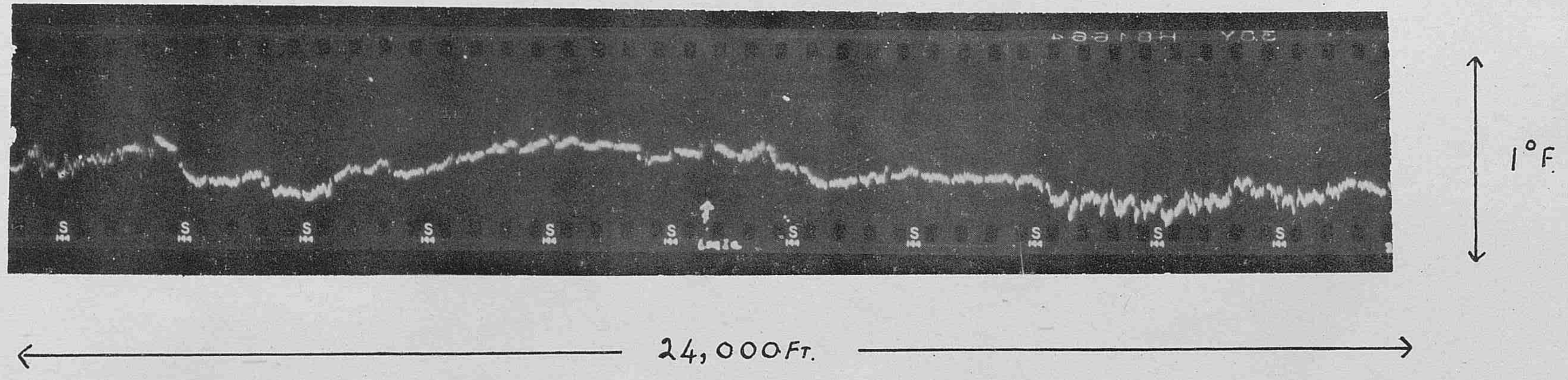


FIG. 11. NOSE OF HASTINGS T.G. 618 SHOWING BOTH SENSITIVE HEADS READY FOR OPERATION.

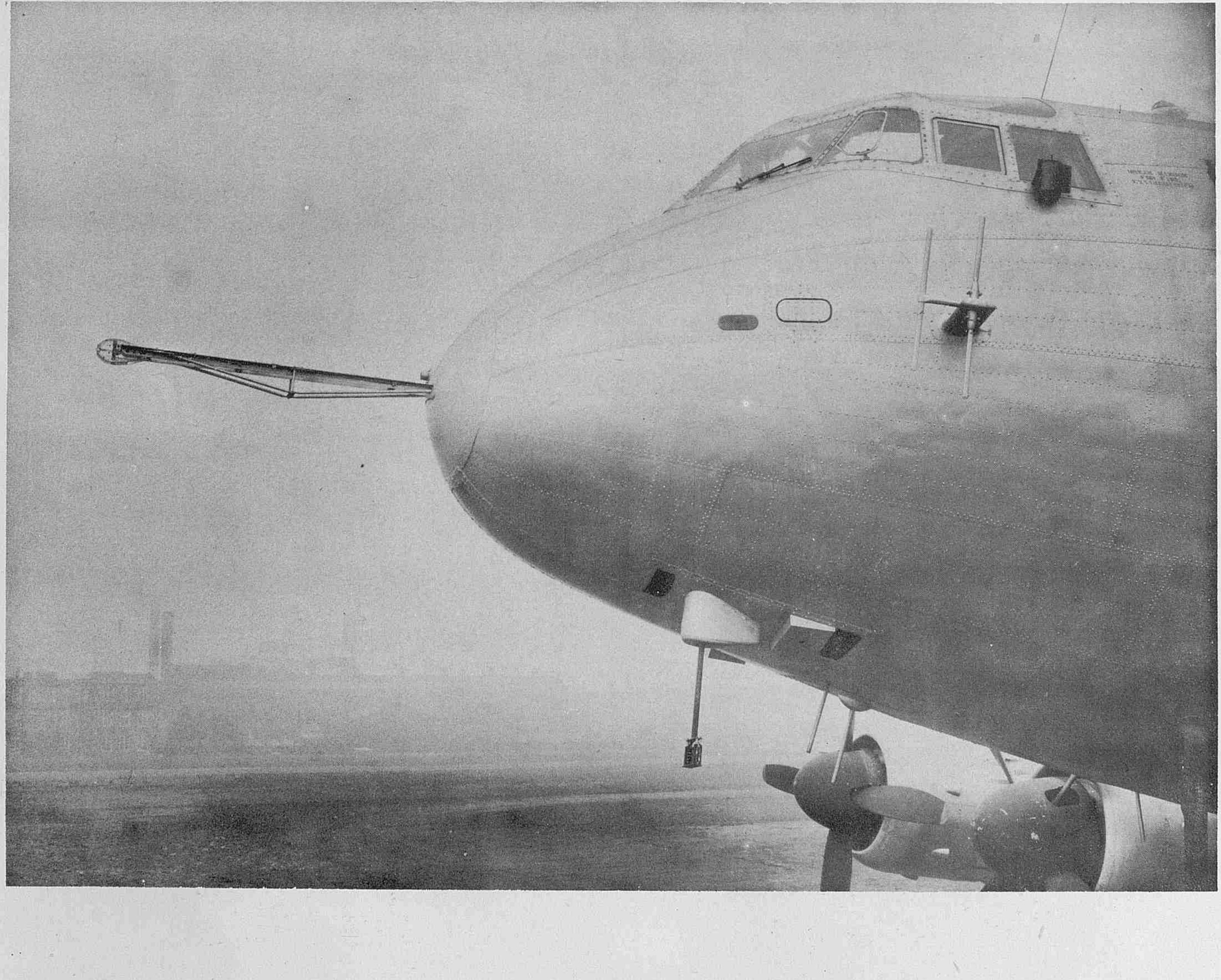
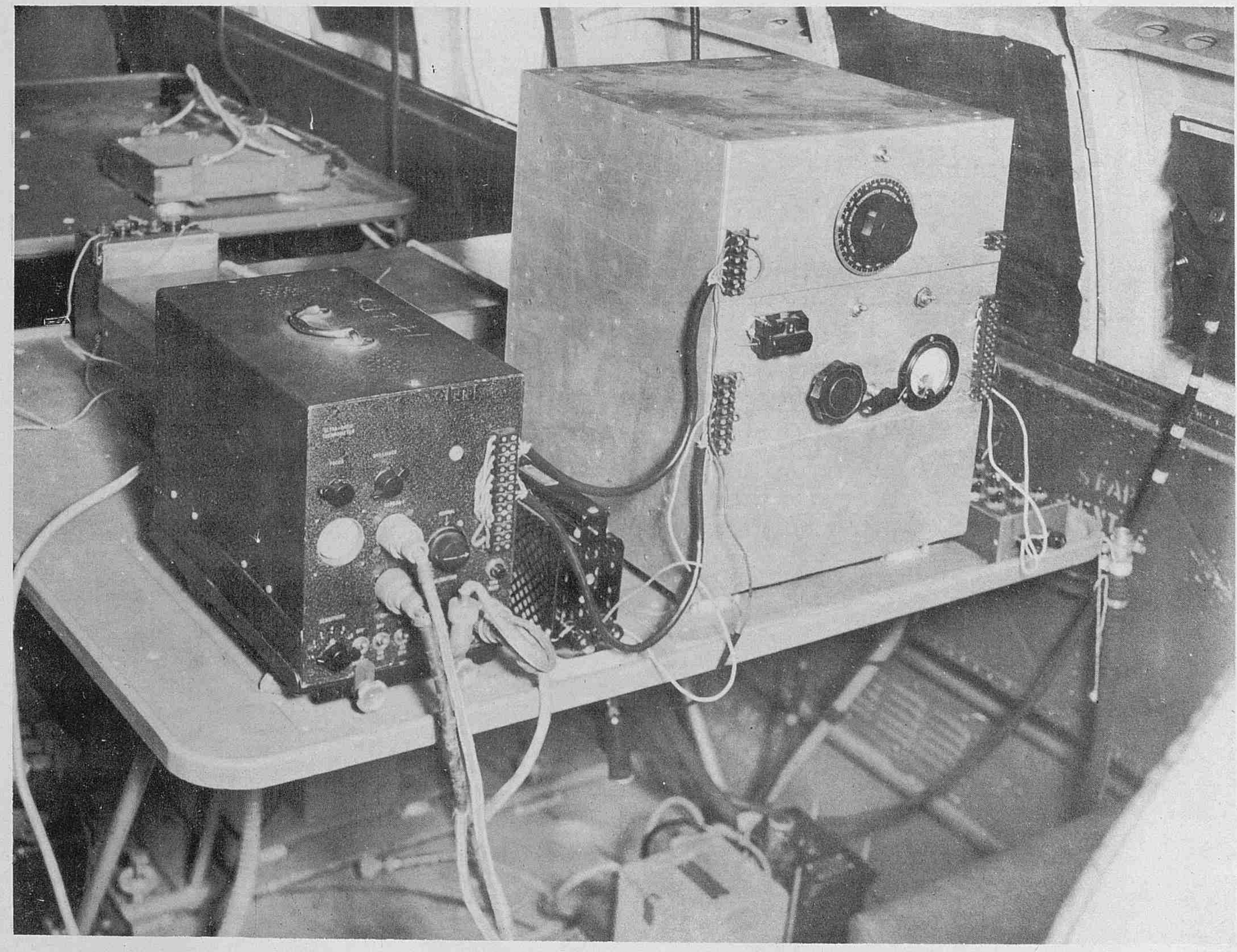


FIG 12. CONTROLLING BRIDGES AND DISPLAY SYSTEM.



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A copy is available in the Library of the Meteorological Office."

M.R.P. 807
S.C.II/147
9 th May, 1953

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## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# The free atmosphere in the vicinity of fronts - analysis of observations by the Meteorological Research Flight (1950-52)

by

J.S. Sawyer (Meteorological Office)

## 1. Introduction

During the period November, 1950 to July, 1952 a series of flights was made by the Meteorological Research Flight the purpose of which was to study the detailed structure of the atmosphere in the neighbourhood of fronts. Twenty-three such flights were made during this period, and all were successfully completed with the exception of one which was cut short by engine trouble. The results of these flights are analysed in the present paper.

The general plan of each of the flights required that the aircraft should make a horizontal flight through the frontal region and return on a reciprocal track at a different level. The track was chosen to be approximately perpendicular to the front. The flight also included spiral climbs or descents at the extremities of the horizontal legs and occasionally at other points. Meteorological observations were made at ½ - minute intervals throughout the horizontal legs and at vertical intervals of 500 or 1000 ft. on the ascents and descents. The aircraft, a Hastings of the Meteorological Research Flight, operated from Farnborough, but the fronts to be investigated were selected at C.F.O. and the general plan of the flight was decided in consultation with the S.Met.O. of the M.R.F. Most of the flights were made in the afternoon to facilitate comparison of the observations with the routine radio-soundings at 1500 G.M.T.

It was decided at the outset that the most useful results would be obtained by exploring a variety of types of front at a variety of levels - thus some fronts were studied only in the lower troposphere and others only between 15,000 and 20,000 ft. It was hoped that the observations might bring out various common properties of all fronts; however, only in their broadest features, such as the existence of a sloping baroclinic zone and of a zone of relatively dry air, do the individual fronts appear to be similar. In studying the results of the flights it is therefore important to remember that none of the 23 fronts which have been explored can be regarded as representative of fronts in general; neither have we examined sufficient fronts of any one type at a particular level (e.g. active warm fronts in the lower troposphere) to

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/Operational

T.I.S
(W1-V1-Gt Br)

G.366364/DOP/4/53/40

Operational limitations on the availability and endurance of both aircraft and crew restricted the number of fronts which could be investigated. In the early stages of the work a desire to make use of every possible occasion resulted in the inclusion of a number of very weak and inactive fronts among those investigated. It was later realised that more readily interpreted results would be obtained if flights were restricted to active fronts, and later fronts were chosen accordingly. The results of all flights are however treated together below.

## 2. Observational Technique

So far as possible the following observations were made and recorded at $\frac{1}{2}$ - minute intervals throughout the horizontal sections of the flights and at a vertical separation of 500 or 1000 ft. on the climbs and descents:-

(a) Temperature

(b) Frost point (or dew-point above 32°F.)

(c) Cloud, precipitation and icing, etc.

(d) Turbulence

(e) Height and airspeed of aircraft.

In addition, winds were obtained by the navigator from the apparent drift of the aircraft between "Gee" fixes. However, these winds did not prove to be sufficiently accurate to extend the knowledge of the wind field which could be derived from the routine radiowind observations. The aircraft observations of wind will not be further discussed.

(a) Temperature Temperatures were read on a flat-plate electrical thermometer mounted beneath the nose of the aircraft. The absolute accuracy of the readings of the thermometer is about ± 0.2°F., and the instrumental lag is small and unlikely to have affected the readings materially in horizontal flight. The thermometer was read to the nearest whole degree on most of the flights, but on some readings to 1/10°F. were recorded. The same correction for the airspeed of the aircraft was applied to observations in cloud and in clear air. This may introduce some uncertainty regarding temperatures in cloud because the air impinging on the thermometer may be compressed under moist-adiabatic conditions instead of the dry-adiabatic conditions applying in clear air.

20 - 12
22 - 0

(b) Frost point or Dew-point Frost points (or dew-points when above 32°F.) were obtained by means of a manually operated Dobson-Brewer frost point hygrometer. The probable error of these observations is about 1°F. at frost points of +60° to 0°F. and 2°-3° at frost points of 0° to -50°F. Records of frost point are not so complete as those of temperature on account of the difficulty of maintaining a long series of readings at only $\frac{1}{2}$ - minute intervals.

(c) Cloud Visual observations of the cloud above and below the aircraft were made and recorded at suitable intervals together with notes on precipitation, icing, etc.

(d) Turbulence The "bumpiness" of the flight was estimated and recorded in terms of "tenths of gravity". The degree of turbulence was recorded at $\frac{1}{2}$ - minute intervals on most of the flights.

(e) Airspeed The airspeed of the aircraft was recorded with a view to its use to assess the direction and magnitude of the vertical movements of the air as was done by Frith (1950).

/3.

## 3. Synoptic analysis of the observations

On any one flight it was possible to explore only a small part of the frontal region. The horizontal extent of the flight, 250 to 300 mi., was usually sufficient to cross the most strongly baroclinic part of the frontal region, but the restriction of the investigation to two levels left a large part of the vertical structure of the front unexplored. In order to complete the picture, and to relate the aircraft's path to the frontal zone, it was necessary to construct an atmospheric cross-section along the flight path on the basis of the routine radiosonde and wind observations. Such a section was constructed for each of the flights. Layers of stable lapse-rate relative to the air above and below were picked out from the radiosonde observations and regarded as defining the "frontal zones"; their boundaries were indicated on the cross-sections. Only the broad features of the soundings were used in placing the frontal zone because it is known that small-scale irregularities of temperature render details of the soundings quite unrepresentative (see Sawyer, 1952).

Comparison between the temperature recorded by the aircraft and temperatures interpolated from the radiosonde cross-section show some systematic differences due to instrumental inaccuracy. On all flights except one the average difference (aircraft minus radiosonde) was negative. The average difference increases from -0.8°F. below 5000 ft. to -2.3°F. above 15,000 ft. Some lag is evident during climbs and descents because the difference aircraft minus radiosonde averages -0.7°F. on climb, - 1.3 on horizontal flight and -2.3 on descent. These discrepancies have been taken into account in analysing the data.

## 4. Thermal structure of the frontal zone

The general trend of temperature as given by the aircraft observations on each of the horizontal flights agrees satisfactorily with the temperature gradient shown on the corresponding cross-section as drawn from radiosonde data alone, but there are appreciable differences in detail. Altogether the flights provided observations along 45 horizontal tracks 29 of which passed through a frontal zone as analysed from the radiosonde observations. The observed profiles of temperature along these 29 flight paths are reproduced in Fig. 1, and the points which have been identified as the limits of the frontal zone have been indicated. In some cases the irregularity of the temperature profile is such that an analysis into a definite frontal zone containing a strong temperature gradient separating two more or less homogeneous air masses represents a considerable idealisation of the observed curve and cannot be carried out objectively; in others the analysis in terms of a frontal zone appears simple and natural. Of the 29 horizontal flights which passed through frontal zones, 20 showed a region of concentration of the horizontal gradient of temperature in close conformity with the frontal zone which was determined from radiosonde data. Such differences in position and intensity as occurred in these cases are little greater than would be expected from the uncertainties in the position of the aircraft and the differences in time between the observations. On the remaining 9 horizontal flights it was difficult to identify the frontal zone satisfactorily.

The most clearly defined frontal zones were found in the upper troposphere in association with cold fronts. The striking example of Nov. 3, 1950, for which profiles at 485 and 550 mb. are included in Fig. 1, was obtained on the first flight of the series. The weather situation is illustrated by the synoptic charts and cross-section in Fig. 2.

An example of the absence of a clearly defined frontal zone is given by the flight of March 19, 1952 which investigated an active warm front off the west of the United Kingdom. At the lower flight level (860 mb.) temperature increased irregularly toward the warm front throughout the flight. At the upper flight level (710 mb.) the main temperature gradient appears to be above the surface warm front (see Fig. 1), but this is difficult to reconcile with the radiosonde observations which indicate the main temperature contrast further east; it also implies a baroclinic zone inclined towards the warm air. The synoptic situation is illustrated in Fig. 3 together with an atmospheric cross-section along the flight path.

In order to present some information on the width of the frontal zones and the temperature change across them, the histograms in Fig. 4 have been constructed. The limits of the frontal zones have been identified on the horizontal temperature profiles as indicated in Fig. 1. For comparison the radiosonde cross-section has been used to make an estimate of the overall change in temperature between the two air masses, and the distance over which a significant temperature gradient extended. This we may call the "frontal region" and histograms concerning its width and temperature contrast are given in the lower diagrams of Fig. 4 for those same occasions and levels for which flights through the frontal zones were made.

The average width of the "frontal region" was about 600 mi. and the average temperature difference across it about 15°F. The horizontal flights show frontal zones with an average width of about 130 mi., that is about a quarter of the total width of the frontal region, and a temperature difference of 9°F., a little more than half of the total air mass contrast. The variation from front to front and the uncertainties of the identification of the frontal boundaries are such that these figures can be little more than a broad indication of the usual structure of a frontal zone, but it is fairly clear that in the free atmosphere zones less than 50 mi. in width are unusual. (On all of the three occasions with frontal zones less than 50 mi. wide the temperature contrast was small, 4°F. in two cases and 6°F. in the other; moreover in two of the cases it is somewhat uncertain whether the aircraft completely crossed the frontal zone.)

Average profiles of temperature across the fronts are given in Fig. 5 based on the aircraft observations. It might appear that the cold fronts were more clearly defined than the warm fronts, but the sample is insufficient to generalise from this data. It is not possible to obtain an entirely satisfactory profile of surface temperature to compare with the profiles of temperature observed by the aircraft. Examination of the surface synoptic charts for the occasions of the flights suggests that the gradient of surface temperature was rather less than in the free atmosphere, and was spread over a distance which also tended to be somewhat less than in the free air.

From the available data it is possible to estimate the slope of the front in three essentially independent ways; these are (a) from the heights of the limits of the frontal zone as given by the radiosonde observations, (b) from the ratio of the vertical thickness of the frontal zone (estimated from the radio-soundings) to its horizontal thickness (determined by the aircraft) and (c) from the relative positions of the frontal zones as determined by the aircraft at the two different flight levels. The values obtained by these three methods are given in the appendix. On only nine occasions did the aircraft cross the frontal zone completely at two levels and method (c) can be applied only in these cases. On one of these occasions (March 19, 1952) the frontal zone is difficult to identify and method (c) gives a slope upward towards the warm air (see also Fig. 3). For the remaining 8 cases the average values of the slope ($\tan \theta$) obtained by the three methods were as follows:-

| | | |
|---|---|---|
| (1) | from radiosonde data alone | .012 |
| (2) | from vertical and horizontal thickness of the frontal zone | .019 |
| (3) | from the positions of the aircraft intersections with the frontal zone. | .019 |

From the figures in the appendix it will be seen that although the three estimates agree as to the order of magnitude of the slopes of the fronts and there is some correlation between them, numerical agreement is poor. Although the extreme estimates are probably in error there is reasonably good evidence that the slopes varied from 1 in 250 to 1 in 30.

Results such as those of the preceding paragraphs can be explained if a sloping frontal zone exists but the position of its boundaries in both horizontal and vertical soundings is subject to considerable uncertainty (say 40 mi. in the horizontal and 2,000 ft. in the vertical). Such uncertainty exists partly because the observations are not synchronous and cannot be positioned with

/sufficient

sufficient accuracy, but also, perhaps more important, because the irregularity of the temperature profile in vertical and horizontal precludes the exact identification of a frontal zone.

## 5. Irregularities of the temperature profile

As has been seen, important difficulties in interpreting the observed temperature profiles arise from the presence of irregularities of temperature which are not representative of the general field of temperature associated with the front. Such irregularities are sufficient to reverse the general horizontal temperature gradient and make it impossible to decide with any certainty where the limits of the frontal zone lie. An example of particularly large irregularities of temperature is provided by the observations at 850 mb. on June 12, 1950 (see Fig. 6). This flight was made through an active cold front and although there is an overall rise of temperature from west to east, it is clearly quite impossible to define a frontal zone. On the other hand, on such temperature profiles as those for Nov. 3, 1951 (see Fig. 1) the irregularities appear only as embroidery upon a clear general trend.

A measure of the magnitude of these irregular temperature fluctuations is given by the standard departure of individual temperature observations from a 5-minute running mean (roughly over 18 mi.). The value of this standard departure for flight in clear air, 0.50°F. is appreciably less than that for cloud 0.70°F. These values are taken from observations for which the mean could be based on observations entirely within either clear air or cloud, and thus refer to points more than 2 minutes flying time from the edge of cloud or precipitation. It is significant that the standard departure, 0.84°F., for the remaining observations which were all near the boundary of cloud or precipitation is greater than that for either clear air or cloud. On most of the flights temperatures were available only to the nearest degree and values of the standard departure less than 0.5°F. are unreliable but it is very unlikely that the differences between clear, cloudy and transitional air arise from an accident of the rounding off procedure. It is possible that the differences are due to a greater random error of temperature measurement in cloud than in clear air, although it seems more likely that the differences are real.

The average magnitude of the temperature irregularities decreases with height above ground, the standard departure from the running mean falling from about 0.8°F. about 5,000 ft. to less than 0.4°F. about 15,000 ft. This marked trend is due partly to the greater proportion of observations in cloud and transitional air at the lower levels and partly to the decrease of the irregularities with height in cloudy and transitional air. There is only a slight tendency for the irregularities of temperature in clear air to decrease with height. These facts can be seen clearly in Fig. 7 in which the standard departure of temperature from the running mean is plotted against the height of the aircraft for observations on individual horizontal legs. Separate diagrams indicate the height variation of temperature departures in cloudy, clear and transitional air.

There was very little difference between the average magnitude of the temperature irregularities in the warm and cold air masses or between these and the frontal zone; the mean values of the standard departure were warm air 0.59°F., frontal zone 0.70°F., cold air 0.69°F. and the differences are not great enough to be regarded as significant.

Although significant irregularities of temperature occur away from cloud and precipitation, there can be little doubt that the largest fluctuations of temperature arise on the borders of cloud and rain from the processes of condensation, evaporation and precipitation. Changes of temperature of 5°F. or more in 2 to 3 minutes (that is in a horizontal distance of 6 to 10 mi.) were observed on several occasions. Temperatures did not consistently rise or fall as the aircraft entered cloud, although on the average the temperature within the cloud appeared to be about 0.9°F. lower than in the adjacent air outside the cloud - a difference which can probably be explained by the fact that the correction applied to the thermometer readings for the compression of the air as it passes over the thermometers takes no account of the cooling effect of the evaporation of cloud droplets. Irregularities of temperature such as those discussed here are not peculiar to frontal regions. Numerous observations of similar irregularities have been made by M.R.F. within individual air masses (see Frith, 1950), and their magnitude probably differs between fronts and air masses only on account of the greater frequency of cloud and precipitation in association with fronts.

## 6. Humidity in frontal regions

Observations of dew-point above 32°F., and frost point below 32°F., were made throughout most of the flights. The frost point (and dew-point) were found to be much more variable than the temperature; fluctuations of the order of 8 or 9°F. over distances of 15 mi. occurred throughout much of the flights. These are some five times as great as the corresponding variations of temperature. Similar irregularities of frost point have been studied by Frith (M.R.P. 402) within individual air masses and are not specifically associated with frontal regions.

The aircraft observations of humidity are not sufficient to establish the complete horizontal and vertical distribution of humidity which accompanied the fronts. However, it was found possible to use the dew-points reported by the routine radio-soundings to prepare a vertical dew-point cross-section through each front. Although such sections cannot show the complex and irregular distribution of water vapour which the aircraft observations demonstrate, nevertheless comparison between the cross-sections and the aircraft observations of humidity shows that the cross-sections are sufficiently accurate to indicate the broad features of the humidity pattern.

A feature, more or less well developed on all of the dew-point cross-sections, was a tongue of relatively dry air extending downward in the vicinity of the front and tilted in the same direction as the front (see Figs. 2 and 3). The dry tongue was less well developed in association with the cold fronts than with the warm fronts; it extended down to an average pressure level of 700 mb. in cold fronts and 800 mb. in warm fronts. In about half of the fronts the driest air was found within the frontal transition zone itself, but there were also occasions when it was found both on the cold and the warm sides of the transition zone. Some striking changes of frost point occurred as the aircraft entered or left the dry tongue; a good example, that of Jan. 4, 1952, is reproduced in Fig. 8. About half of the flights showed a sharp transition from moist to dry air and the change of frost point averaged about 37°F. in 35 mi. on these flights. Some of the sharpest transitions gave changes of more than 40°F. in 20 mi.

Average profiles of frost point across the frontal zones have been computed from the aircraft observations for cold fronts and warm fronts; these are reproduced in Fig. 5. The results depend slightly on the particular technique of averaging, and the method adopted here is explained on the diagram. However the essential features are not in doubt. At warm fronts the profile shows a steady change from moist warm air to dry cold air over a distance of about 300 mi., whereas for cold fronts the average profile indicates a dry transition region between two relatively moist air masses - the dry zone is about 100 mi. wide. The cold front profile confirms the existence of the dry tongue described in the preceding paragraph; the reason for its appearance only on the extreme right of the average profile for warm fronts is to be found in the variation in the position of the dry tongue relative to the front, together with the dryness of the cold air mass.

The very rapid increase of frost point as the cloud is approached may be noted in Fig. 8; this is not uncommon. Table I gives the average difference between the temperature at the edge of the cloud and the frost point at points within the clear air distant from the cloud by $\frac{1}{2}$, 1, $1\frac{1}{2}$ etc. minutes flying time. It shows that the humidity falls off rapidly away from the cloud so that at a distance of less than 10 mi. the average depression of the frost point is 10°F.; it changes only slowly with further distance from the cloud. From these results it seems improbable that, in the analysis of free air humidity observations, extrapolation of the humidity and temperature observations to saturation can accurately predict the position of cloud masses.

Table I Average difference between frost point at various distances from the edge of a cloud and the temperature at the cloud boundary

| Flying time from cloud edge (min.) | | $\frac{1}{2}$ | 1 | $1\frac{1}{2}$ | 2 | $2\frac{1}{2}$ | 3 | $3\frac{1}{2}$ | 4 | $4\frac{1}{2}$ | 5 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Approx. distance from cloud edge (mi.) | | 2 | 4 | 5 | 7 | 9 | 11 | 13 | 15 | 16 | 18 |
| Temperature at cloud edge minus frost point o | °F | 5.9 | 7.8 | 8.3 | 9.0 | 10.6 | 13.0 | 13.0 | 12.0 | 11.6 | 13.3 |
| | °C | 3.3 | 4.3 | 4.6 | 5.0 | 5.9 | 7.2 | 7.2 | 6.7 | 6.5 | 7.4 |
| Temperature at cloud edge minus temperature | °F | -0.3 | -0.6 | -0.6 | -0.7 | -0.8 | -0.8 | -0.8 | -0.7 | -0.7 | -0.8 |
| | °C | 0.1 | 0.3 | -0.3 | -0.4 | -0.4 | -0.4 | -0.4 | -0.4 | -0.4 | -0.4 |

An interesting question concerns the source of the dry air which occurs in the "dry tongues". In order to study this, trajectories of the air within the dry tongue, and of the adjacent moister air were drawn on the basis of routine contour charts for appropriate pressure levels. Although the trajectories are necessarily rather uncertain, the analysis shows that it is reasonable to attribute the humidity differences primarily to large-scale horizontal air movements. The dry air accompanying a warm front seems to have been derived from the subsided air of the preceding ridge and may not have undergone any recent large vertical movement when found in or just ahead of the frontal zone. A typical example is that of Jan. 4, 1952 for which three representative 700-mb. trajectories are shown in Fig. 9 terminating at the points indicated in the cross-section; aircraft observations of temperature and frost point are given in Fig. 8. It is clear that air with widely different histories entered into the frontal region. The air within the dry tongue was off the west coast of Scotland 24 hours earlier, close to the surface ridge of high pressure. The Stornoway radiosonde showed that the air in that region was already notably dry, the dew-point being reported as -22°F. at 700 mb. and -41°F. at 650 mb. The temperature is somewhat uncertain because the trajectory passed some 150 mi. west of Stornoway. It is doubtful if the Stornoway 700 mb. temperature at 1500 G.M.T. can be regarded as applicable at this point because between 1500 and 2100 G.M.T. the temperature increased from 3°F. to 9°F. This was probably largely due to subsidence, but if it was due to advection then temperature must have increased rapidly to the west of Stornoway. In either case it is clear that the subsidence which effectively dried the air took place in the ridge of high pressure and some 18 hours or more before it was sampled by the M.R.F. aircraft.

A particularly interesting case is that of the cold front of Nov. 3, 1950, in which very dry air was found in the transition zone. Relevant trajectories for air moving at the 500 mb. level are shown in Fig. 10. There was undoubtedly a very marked confluence of air from widely separated areas, and the air which was in or close to the frontal zone at the time of the flight was not associated with any frontal zone 24 hours earlier. The exact source of the air which lay in the transition zone at 1500 G.M.T. Nov. 3, 1950 is uncertain, but it appears most likely that it was off the coast of NW. Scotland. The air in this region was cold and the tropopause low (373 mb. at Stornoway at 1500 G.M.T. Nov. 2,) and if air moved adiabatically to the frontal zone over France where the temperature was between -8°F. and 0°F. at 500 mb. then it would have had to descend by about 140 mb. so that stratospheric air was brought down to the 500 mb. level. (Such vertical motion casts doubt on the value of a trajectory based on the 500 mb. level but if a trajectory is interpolated between 500 and 300 mb. making allowance for the vertical motion little material difference is introduced). Trajectories of air on the warm and cold sides of the frontal zone indicate little change of temperature at 500 mb., and the greatest descent of the air appears to have been confined to a limited band of air from northwest of Scotland which was drawn out into a narrow filament as it passed into the entrance region of the jet stream. Thus the very dry transitional air observed on Nov. 3, 1950 may have been derived from the stratosphere within the preceding 24 hours.

On only two other occasions was the upper part of a cold front investigated (viz on Apr. 27, 1951 and Aug. 1, 1951). Both showed very dry air in or near the frontal zone. However trajectories drawn to show the history of the air at the 500 mb. level indicate that warm, cold and transitional air had moved horizontally on roughly parallel trajectories through the jet stream system during the last 24 hours. Whether any process similar to that of Nov. 3, 1950 had taken place at an earlier stage cannot be established owing to lack of observations in the relevant areas.

## 7. Cloud observations in association with fronts

From the observations of cloud made by the M.R.F. aircraft at two different levels together with the observations of surface reporting stations, it was possible to form a reasonably coherent picture of the distribution of cloud accompanying the fronts through which the flights were made. Figs. 3, 11 and 12 illustrate the type of cross-section which it was possible to construct.

The distribution of cloud varied widely between individual fronts but the fronts which were investigated may be grouped as in Table II to demonstrate the main features of the cloud structure.

Table II Analysis of cloud structure at fronts from the observations of M.R.F. aircraft

| | No. of cases |
|---|---|
| Main cloud mass in the warm air mass with only a relatively small extension into the transition zone | 9 |
| Main cloud mass partly in warm air mass and partly in the transition zone | 7 |
| Frontal cloud only in or near boundaries of transition zone. (These were all weak fronts and cloud was patchy and not extensive) | 4 |
| Frontal cloud consisted of a mass of convective cloud in cold air and transition zone - front thermally ill-defined | 1 |
| Solid mass of cloud in cold air mass (the cloud was probably associated with an old occlusion ahead of the main thermal front - see further notes in section 10(a)) | 1 |

The amount and extent of cloud encountered varied considerably. Several fronts were crossed without the aircraft flying through any cloud at the selected level; at the other extreme, a horizontal flight of almost 300 mi. was made entirely in cloud (the case is further discussed in section 10 (d)). An indication of the variation of the amount of cloud with height is given by Table III which shows the proportion of total duration of the horizontal flights which took place in cloud. As each front was only investigated at two levels and those not often widely separated, the figures in Table III are no doubt strongly influenced by the accidental choice of fronts and flight levels. However the fairly uniform distribution of cloud at all levels up to 15,000 ft. and the relatively smaller amounts above 15,000 ft. conform with expectation and are probably real.

Table III Duration of flight in cloud and clear air at various levels during series of flights

| Height range | Duration of flight in cloud | Duration of flight in clear air | Proportion of flight in cloud |
|---|---|---|---|
| | min. | min. | % |
| Above 15,000 ft. | 139 | 627 | 20 |
| 10,000 to 15,000 ft. | 386 | 868 | 31 |
| 5,000 to 10,000 ft. | 193 | 367 | 34 |
| Below 5,000 ft. | 197 | 415 | 32 |
| All levels | 915 | 2277 | 29 |

When viewed from below the frontal cloud masses were generally reported as being of layer type; altostratus, nimbostratus or altocumulus. Reports of the cloud from above were not so numerous but were again mainly of layer type. Reports of layer cloud predominate in the cold fronts as well as in the warm fronts, but convective cloud is found associated with the frontal cloud mass in all of the five cold fronts for which information is available. Convective cloud was reported in association with the frontal cloud of 6 out of the 17 warm and quasi-stationary fronts; most of these reports are consistent with the unstable or near neutral lapse rates in the regions concerned.

The two cross-sections of cloud structure in Figs. 11 and 12 have been reproduced to illustrate some of the more common features of the cloud distribution. It would be wrong to regard these sections as typical of the flights as a whole, but some of their features occurred several times in the small sample which was available, and these diagrams may perhaps be regarded as illustrating common types of cloud structure at the more active warm and cold fronts.

The section for March 6, 1952 (Fig. 11) shows a solid mass of cloud some 200 mi. broad at most levels and extending upward well above the higher flight level (650 mb.). The upper limit of the cloud is uncertain, but has been inserted on the basis of the humidities reported on the radiosonde soundings. A noteworthy feature, which also occurs at other warm fronts, is that the general inclination of the cloud mass is steeper than that of the front. The dashed lines of Fig. 11 indicate the wet-bulb potential temperature required for saturation at the observed temperature. Since the wet-bulb potential temperature in the upper part of the cloud lies mainly between 52 and 58°F., the air comprising this part of the cloud may have been derived from the lower layers of the warm air mass. The lower part of the cloud lying within the frontal transition zone has a lower wet-bulb potential temperature. If it has been derived from the warm air mass it must have been subject to considerable cooling by precipitation. It is more probable that this cloud has been formed in cold or transitional air which has been moistened by precipitation.

Noteworthy features of the cloud structure at the cold front of April 27, 1951 (Fig. 12) are the narrow area of large convection cloud in the warm air mass close to the surface front, and the absence of cloud in the frontal zone in the upper troposphere. Only five other cold fronts were investigated (and for one the cloud observations are incomplete). However, the frontal zone was free of cloud in the upper troposphere in all six fronts, and a rather narrow band of cloud with considerable vertical development occurred in the warm air above the surface front in four out of the five fronts for which this region was explored. The solid mass of cloud which appears in Fig. 12 in the transition zone below 650 mb. was found only in this case, but this front was the most active of the series and such structure may be more common than its occurrence once in five fronts might suggest. On this occasion the aircraft passed through the top of the cloud mass which lay in the frontal zone and reported marked cumuliform structure. This is surprising in view of the likelihood of a stable vertical lapse rate in the frontal zone (confirmed by the Trappes radiosonde). Possibly, however, the vertical convection currents did not extend through any considerable depth.

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## 8. Turbulence in association with fronts

The average level of turbulence recorded in the course of the frontal flights was slightly greater in cloud than in clear air and significantly greater still near the boundaries of cloud and precipitation. Frequencies of various levels of turbulence over 5-minute intervals entirely in cloud, entirely in clear air and partly in cloud or precipitation are given in Fig. 13. The difference of the average level of turbulence is very slight between the warm, cold and transitional air masses although it is probably significant that the lowest average is found in the cold air where less cloud is encountered. The turbulence tends to be somewhat greater in the first 5,000 ft. above the ground but no variation with height above 5,000 ft. was noted.

Probably of greater interest is the occurrence of bumps of 0.3g or greater. Bannon (1951) quotes 0.4g as the lower limit of heavy turbulence; thus by considering bumps of 0.3g and more, we shall include some "moderate" turbulence along with the severecases. Even so, gusts of 0.3g were encountered in less than 5% of the 5-minute intervals into which the flights were divided for this analysis. On the other hand one or more gusts of this magnitude were encountered somewhere on most of the flights.

Gusts of 0.3g or more were encountered on 20 of the horizontal legs. Only two of these occasions occurred in clear air without cloud or rain in the vicinity. The cause of turbulence on these two occasions is obscure because the wind shear does not appear to have been unusually great, nor could the Richardson number have been much lower than normal - certainly not as low as one. On one of these occasions (Nov. 3, 1950, 485 mb.) the bumpiness was found above the upper boundary of a well defined cold front (see Fig. 2); the other observation of clear air bumpiness of 0.3g occurred on May 23, 1951 at 500 mb. close to the lower boundary of a warm front. On the first occasion the bumpiness was very limited in extent; the extent on the second occasion is unknown because the aircraft descended out of the area.

Of the remaining 18 occasions of gusts of 0.3g or more, 13 were associated with reports of cumulus cloud, thunderstorms or lapse rates in excess of the saturated adiabatic, and there seems little doubt that these arose from ordinary convective vertical motion. Two cases of 0.3g bumps occurred in cloud with a lapse rate quite stable for moist air; however, they were both associated with a strong vertical wind shear and a Richardson number probably less than one. Three occasions were accompanied by lapse rates close to the saturated adiabatic. Two of these had wind shears of 6 kt./Km. and 24 kt./Km. respectively and the Richardson number must have been less than unity; the wind shear in the third case is uncertain but was up to 5 kt./Km. over neighbouring areas. The situations in which gusts of 0.3g or more were found are summarised in Table IV.

Table IV Analysis of situations in which turbulence of 0.3g or more was observed

| Situation | No. of cases |
|---|---|
| Associated with convective phenomena, cumulus, thunder etc. | 13 * |
| In cloud with lapse rate near saturated adiabatic and strong wind shear | 3 |
| In cloud with stable lapse rate and strong wind shear | 2 |
| In clear air not close to cloud or precipitation (no obvious explanation) | 2 |
| Associated with lenticular cloud | 1 * |

* Note On one horizontal flight two distinct occurrences of turbulence occurred; one occurred in irregular frontal cloud with lapse rate near saturated adiabatic and is included under convective phenomena, and the other occurred just below lenticular cloud.

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## 9. Airspeed

A visual comparison between the plots of airspeed and those of temperature showed that if any correlation existed between them it would be small; moreover no consistent behaviour of the airspeed in relation to the frontal boundaries could be detected. It was therefore thought unlikely that any statistical study of the airspeed variations could give results which would have any significant bearing on frontal structure and no such analysis was attempted.

## 10. Notes on some individual fronts

### (a) June 26, 1951

On June 26, 1951 (see Fig. 14) a flight was made primarily to investigate the warm front to the southwest of the British Isles. Rain was falling over central England but this was more closely associated with an old depression over the North Sea than with the approaching front. It is not easy to place the frontal zone in the free atmosphere but the only sloping baroclinic zone which can be identified appears to have been situated to the west of and above the area investigated by the aircraft (see Fig. 14). The rain over England was falling from an extensive mass of cloud lying in a baroclinic region within the cold air; at the flight levels (550 and 650 mb.) the cloud was rather tenuous and flying conditions notably smooth. The moisture necessary to form this mass of cloud and precipitation was certainly derived from the old depression over the North Sea, and the old occlusion could have entered into the area. However, there is no indication on the vertical cross-sections of any occluded warm air mass, and it seems more appropriate to regard the cloud and precipitation as due to "development" taking place in the general baroclinic zone and producing the most significant "weather" where the air was moist and easily saturated by lifting.

### (b) August 17, 1951

On August 17, 1951 the structure of an old occlusion was investigated and the relevant synoptic charts and cross-sections are reproduced in Fig. 15. Noteworthy features are:-

(i) the shallow nature of the frontal structure

(ii) the warm dry air above the inversion behind the front

(iii) the relatively shallow layer of moist air which produced the cloud masses. (On account of its wet-bulb potential temperature it appears to have been derived from the lowest 4,000 ft. of the air behind the front)

(iv) the unstable character of the cloud mass and the associated irregular horizontal profile of temperature (see Fig. 1)

(v) the persistent bumpiness on all sections of the flight except in the warm dry air above the inversion to the west of the front. Bumps reached 0.3g occasionally at all flight levels.

### (c) August 23, 1951

The flight of August 23, 1951 is interesting as showing that the structure of a warm front may be weak and ill-defined up to within 300 mi. of the centre of a depression. Fig. 16 shows the synoptic situation and cross-section through the front. The frontal surface is difficult to place on the cross-section, and the cloud appears to lie mainly in two broken layers at the base of the two sloping stable layers of the atmosphere. The existence of somewhat thicker cloud layers above the surface front is also noteworthy.

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### (d) November 5, 1951

The flight of November 5, 1951 demonstrated the existence of a very broad and steeply sloping baroclinic zone in association with a slow-moving front. The cloud must have been formed by vertical motion in the baroclinic zone; no material part of the moisture can have been supplied by precipitation from the overlying warm air because the slope of the front was so steep (1 in 30) that a large part of the transition zone was not vertically beneath any part of the warm air. The broad belt of cloud and precipitation is interesting.

## Conclusions

(a) The present series of flights indicates the great complexity of the atmosphere in frontal regions; no idealised frontal model can adequately represent an individual front.

(b) The flights confirm that fronts are associated with a sloping baroclinic zone and that about half of the total air mass contrast of temperature is concentrated in a region with a width usually of 100-200 mi.

(c) Small-scale irregularities of temperature of the order of 1°F. are sufficient to prevent any clear delimitation of a frontal zone by distinct boundaries except in association with the upper part of some cold fronts where the air is usually cloud-free.

(d) The largest irregularities of temperature are associated with the physical processes of condensation, precipitation and evaporation, and they occur near the boundaries of cloud and precipitation. The most striking gradients of temperature arise from these causes and reach 5°F. in less than 10 mi.

(e) Some very dry air is usually found in frontal regions in close juxtaposition to cloud and precipitation. Its presence can be adequately explained by large scale movements of the air which are primarily horizontal; the air has been dried by subsidence but this may have been 24 hours or more before it was observed in the frontal region. The presence of dry air is no indication of descending motion in the front, nor of the absence of adequate moisture for frontal rain. Thus reported humidity on individual radio-soundings cannot provide a forecaster with any reliable indication of the activity of a front.

(f) Cloud structure of fronts is very variable. The main frontal cloud masses are usually formed within the warm air mass, and extend into the transition zone by the evaporation and recondensation of rain or snow. However, observations on June 26, 1951 and Nov. 5, 1951 suggest that frontal cloud can also form in the cold air or transition zone, if adequate moisture is available there.

(g) Fronts are not ordinarily regions of particularly marked turbulence. Most of the moderate or severe turbulence is associated with the ordinary processes of convection, (cumulus cloud, thunderstorms etc). Some cases were also observed in association with strong vertical wind shear and two occurrences of 0.3g bumps in clear air remain unexplained.

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## Appendix. - List of fronts explored by M.R.F. Aircraft.

| Date | Type | Pressure level (mb.) | Did flight intersect frontal zone? | Width of frontal zone (mi.) | $\Delta T$ | Slope of front | | |
|---|---|---|---|---|---|---|---|---|
| | | | | | | Based on r/s data | $\frac{\Delta z}{\Delta x}$ | Direct observations |
| 3 Nov. 1950 | cold | 485 | Yes | 58 | 18 | .014 | .017 | .015 |
| | | 550 | Yes | 79 | 16 | .014 | .010 | |
| 27 Apr. 1951 | cold | 500 | Yes | 159 | 15 | .007 | .006 | .010 |
| | | 650 | Yes | 195 | 14 | .012 | .008 | |
| 23 May. 1951 | warm | 500 | Yes | 187 | 8 | .016 | .017 | .020 |
| | | 650 | Yes | 231 | 7 | .006 | .008 | |
| 12 June. 1951 | cold | 650 | No | - | - | - | - | - |
| | | 850 | No | - | - | - | - | - |
| 13 June. 1951 | warm | 550 | No | - | - | - | - | - |
| | | 650 | Yes | 107 | 8 | .006 | .014 | - |
| 26 June. 1951 | warm | 550 | No | - | - | - | - | |
| | | 650 | No | - | - | - | - | |
| 27 June. 1951 | warm | Flight abandoned because of engine trouble. | | | | | | |
| 9 July. 1951 | warm | 550 | No | - | - | - | - | |
| | | 700 | Yes | 99 | 5 | .004 | .004 | |
| 1 Aug. 1951 | cold | 500 | Yes | 55 | 4 | .024 | .024 | .032 |
| | | 650 | Yes | 36 | 4 | .024 | .037 | |
| 17 Aug. 1951 | occ. | 650 | No | - | - | - | - | |
| | | 800 | Yes | 220 | 10 | .007 | .002 | |
| | | 850 | No | - | - | - | - | |
| 21 Aug. 1951 | warm | 650 | No | - | - | - | - | |
| | | 850 | Yes | 144 | 13 | .008 | .003 | |
| 23 Aug. 1951 | warm | 650 | No | - | - | - | - | |
| | | 850 | Yes | 62 | 4 | .002 | .006 | |
| 3 Sept. 1951 | warm | 600 | Yes | 86 | 4 | .012 | .012 | .006 |
| | | 800 | Yes | 16 | 6 | .014 | .048 | |
| 5 Nov. 1951 | q.s. | 600 | Yes | 196 | 9 | .022 | .019 | .028 |
| | | 850 | Yes | 199 | 13 | .013 | .019 | |
| 14 Nov. 1951 | warm | 450 | Yes | 111 | 8 | .007 | .017 | .007 |
| | | 550 | Yes | 46 | 4 | .010 | .047 | |
| 11 Dec. 1951 | warm | 600 | No | - | - | - | - | |
| | | 700 | No | - | - | - | - | |
| 4 Jan. 1951 | warm | 700 | Yes | 110 | 8 | .005 | .013 | .006 |
| | | 800 | Yes | 81 | 8 | .004 | .013 | |
| 6 Mar. 1952 | warm | 650 | Yes | 212 | 10 | .006 | .007 | |
| | | 870 | No | - | - | - | - | |
| 7 Mar. 1952 | warm | 670 | No | - | - | - | - | |
| | | 870 | Yes | 149 | 7 | .006 | .010 | |
| 19 Mar. 1952 | warm | 710 | Yes | 93 | 6 | .020 | .027 | .008 |
| | | 860 | Yes | 258 | 7 | .008 | .010 | |
| 25 Mar. 1952 | cold | 500 | No | - | - | - | - | |
| | | 740 | Yes | 121 | 11 | .017 | .013 | |
| 28 May. 1952 | cold | 560 | No | - | - | - | - | |
| | | 940 | Yes | 55 | 6 | .008 | .038 | |
| 3 July. 1952 | cold | 500 | Yes | 173 | 12 | .013 | .011 | .050 |
| | | 650 | Yes | 166 | 9 | .010 | .009 | |

°F.

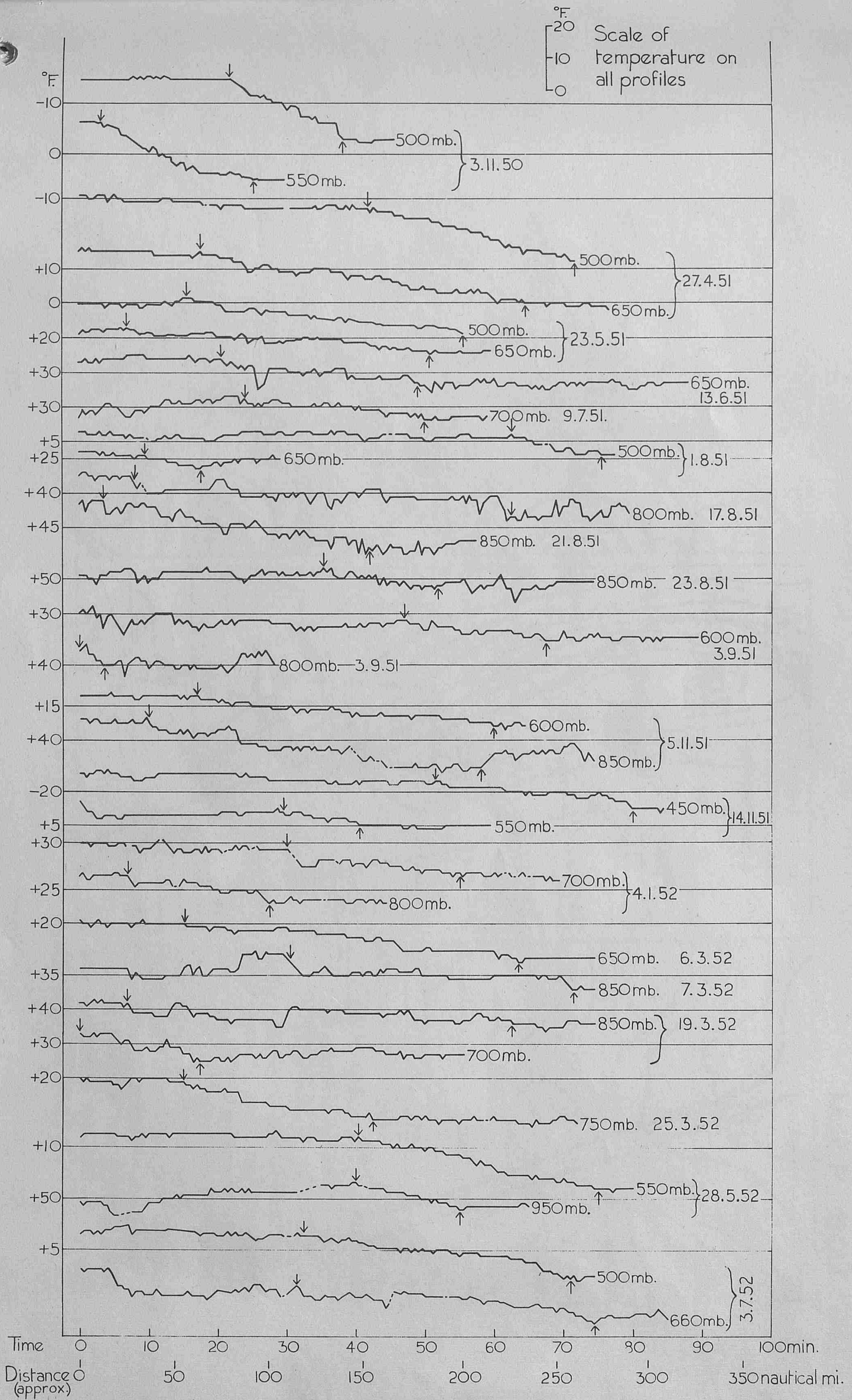

Fig. 1 Profiles of temperature observed by aircraft on horizontal flight through frontal zones
Boundaries of frontal zone as used in analysis of data are indicated thus: ↓ ↑
All profiles are arranged with the warm air mass to the left

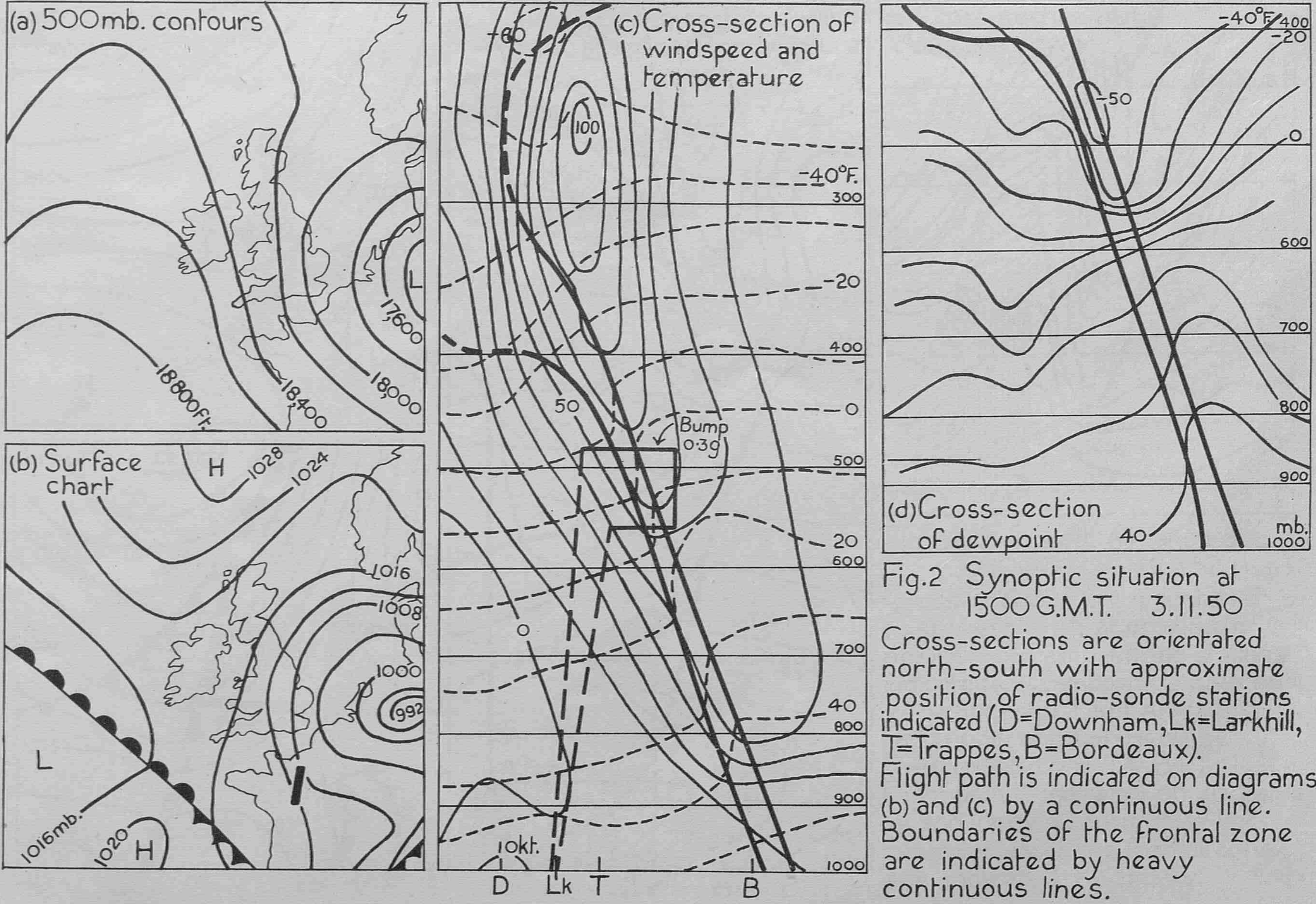

Fig.2 Synoptic situation at 1500 G.M.T. 3.11.50

Cross-sections are orientated north-south with approximate position of radio-sonde stations indicated (D=Downham, Lk=Larkhill, T=Trappes, B=Bordeaux).
Flight path is indicated on diagrams (b) and (c) by a continuous line.
Boundaries of the frontal zone are indicated by heavy continuous lines.

60

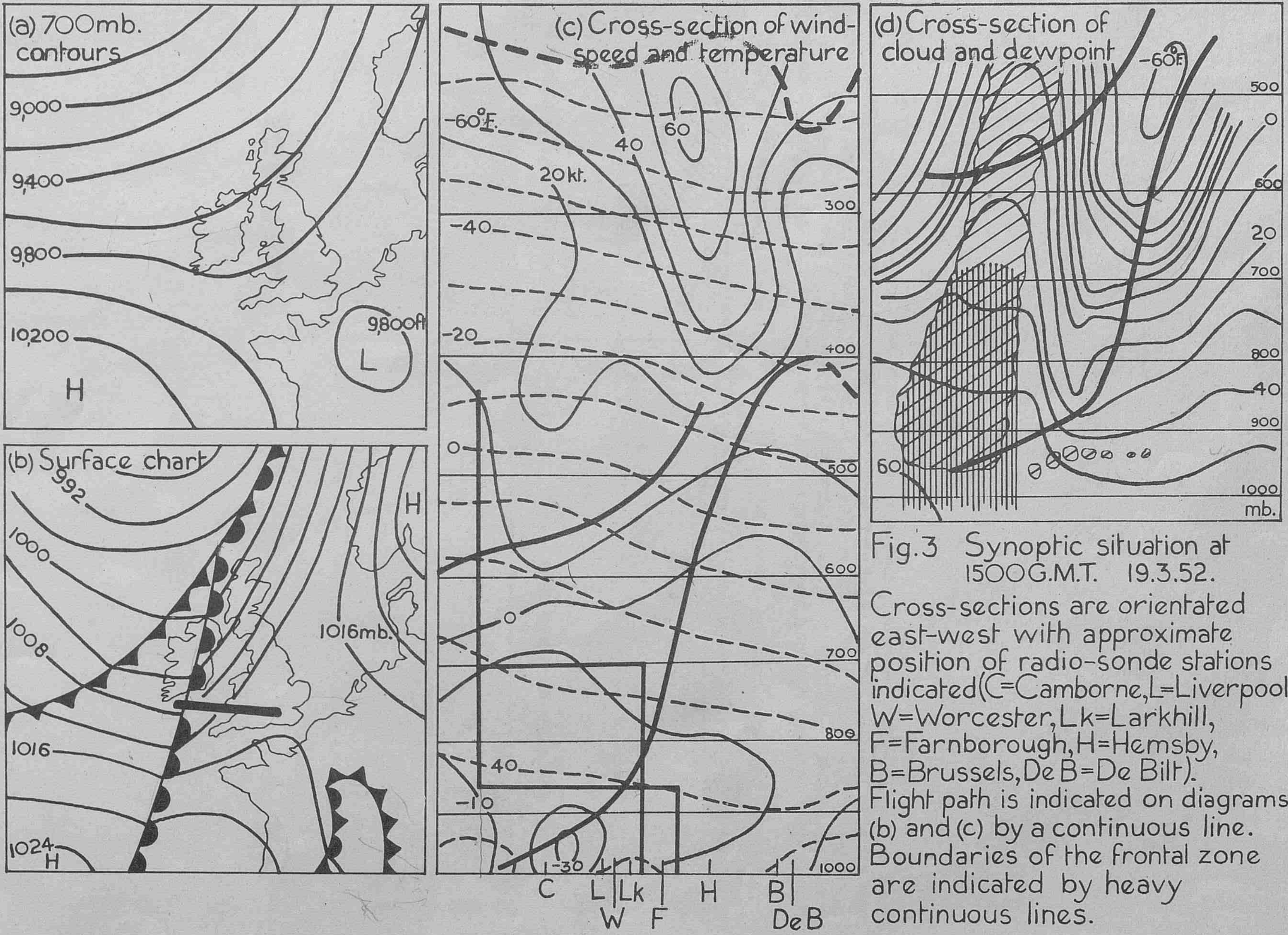
(a) 700mb. contours
9,000
9,400
9,800
10,200
H
9,800ft
L
(b) Surface chart
992
1000
1008
1016
1024
H
1016mb.
H
(c) Cross-section of wind-speed and temperature
-60°F.
20kt.
40
60
-40
-20
0
0
40
-10
-30
300
400
500
600
700
800
1000
C
L
W
Lk
F
H
B
DeB
(d) Cross-section of cloud and dewpoint
-60°F.
0
20
40
60
500
600
700
800
900
1000
mb.

Fig. 3 Synoptic situation at 1500G.M.T. 19.3.52.

Cross-sections are orientated east-west with approximate position of radio-sonde stations indicated (C=Camborne, L=Liverpool, W=Worcester, Lk=Larkhill, F=Farnborough, H=Hemsby, B=Brussels, DeB=De Bilt). Flight path is indicated on diagrams (b) and (c) by a continuous line. Boundaries of the frontal zone are indicated by heavy continuous lines.

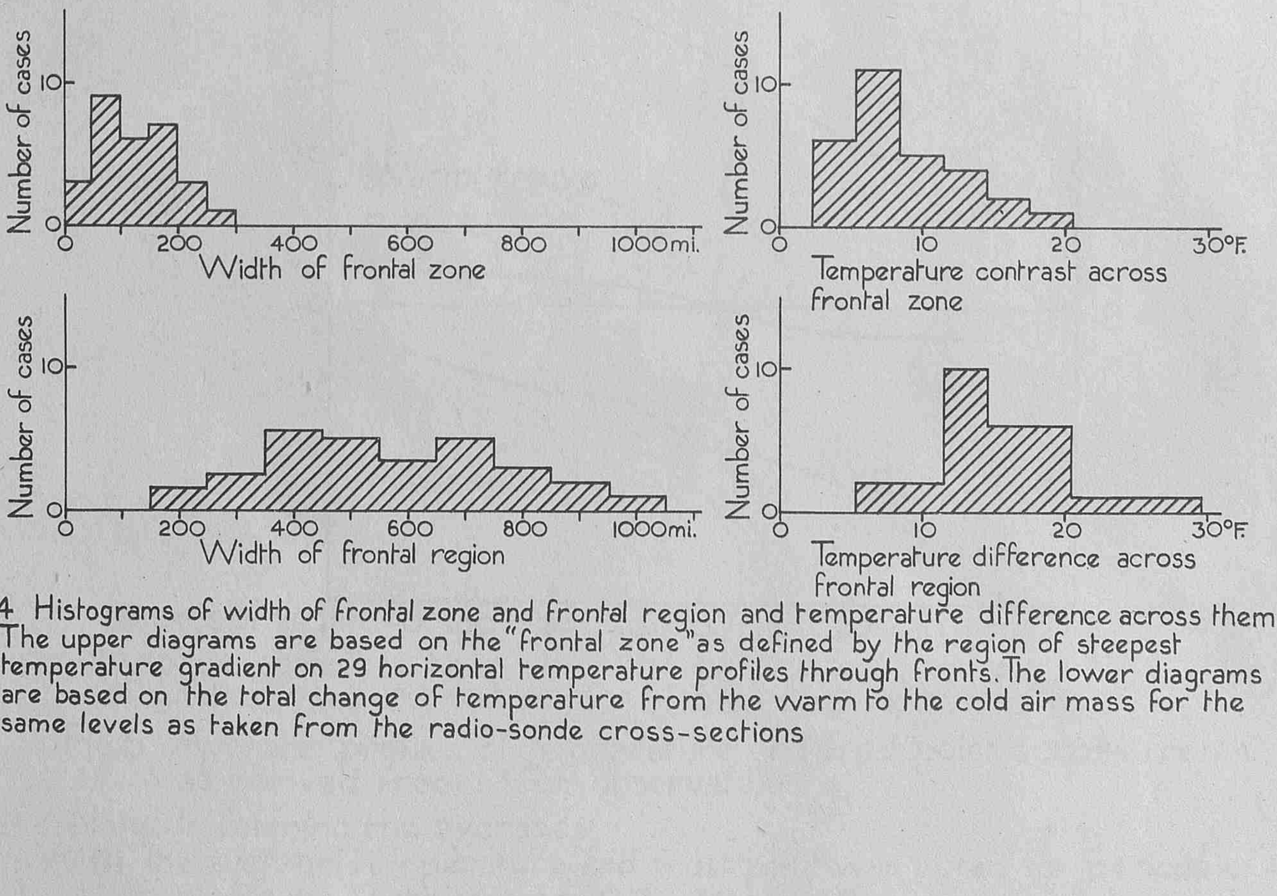

Fig. 4 Histograms of width of frontal zone and frontal region and temperature difference across them. The upper diagrams are based on the "frontal zone" as defined by the region of steepest temperature gradient on 29 horizontal temperature profiles through fronts. The lower diagrams are based on the total change of temperature from the warm to the cold air mass for the same levels as taken from the radio-sonde cross-sections

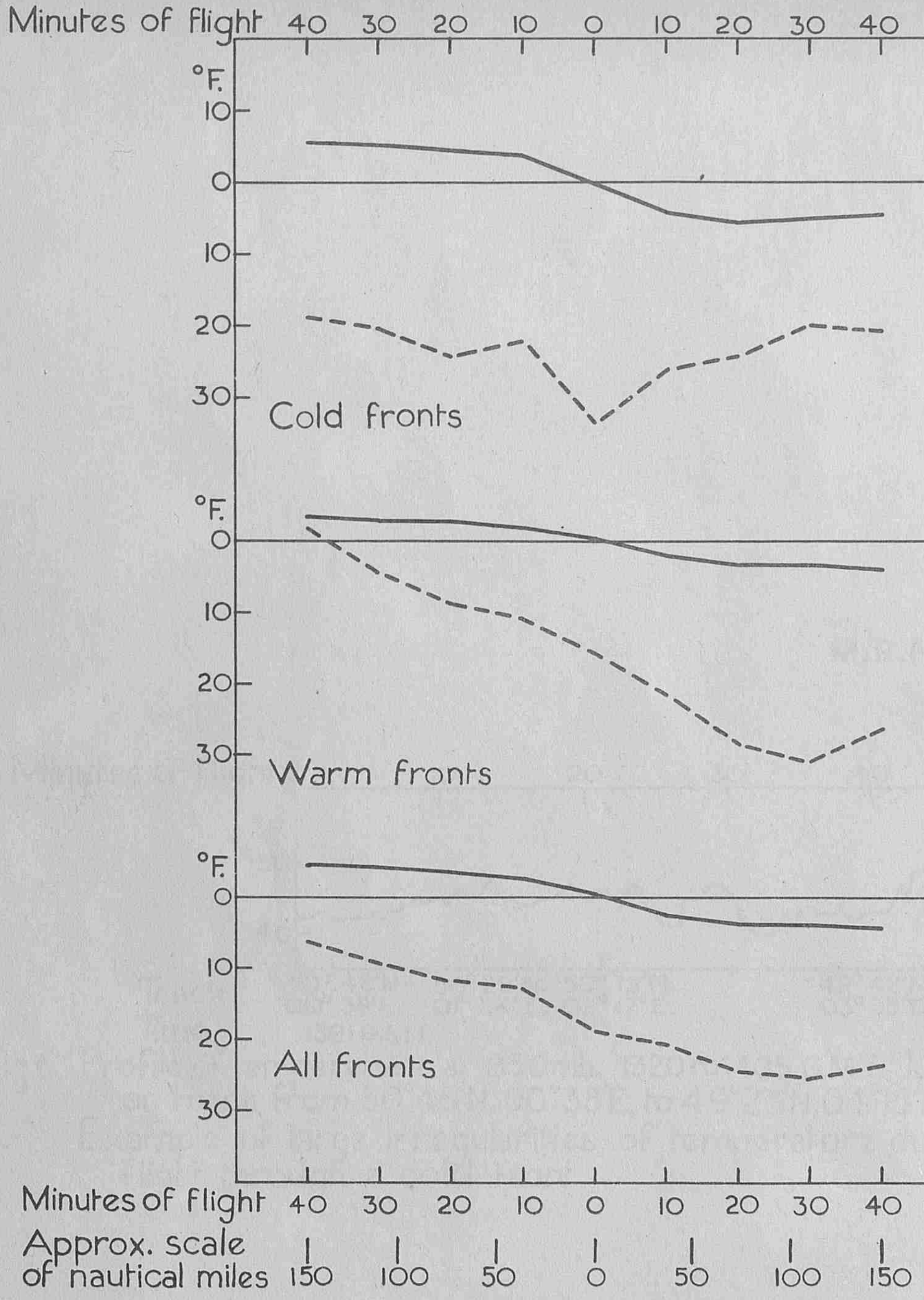

Fig.5 Average profiles of temperature and frost point across fronts as derived from M.R.F. observations

Note: In forming the averages,

(1) The average temperature and frost point was noted for periods of 10 minutes flight centred 10, 20, 30 and 40 minutes from the centre of the transition zone.

(2) The difference of the average temperature from the average for the interval in the centre of the transition zone was calculated.

(3) The mean temperature profile was plotted from the mean value of the differences obtained in stage (2).

(4) Average values of the depression of the frost point were calculated for the intervals at distances 10, 20, 30 and 40 minutes from the centre of the transition zone.

(5) The mean frost point profile was plotted by subtracting these average frost point depressions from the corresponding temperatures of the mean temperature profile

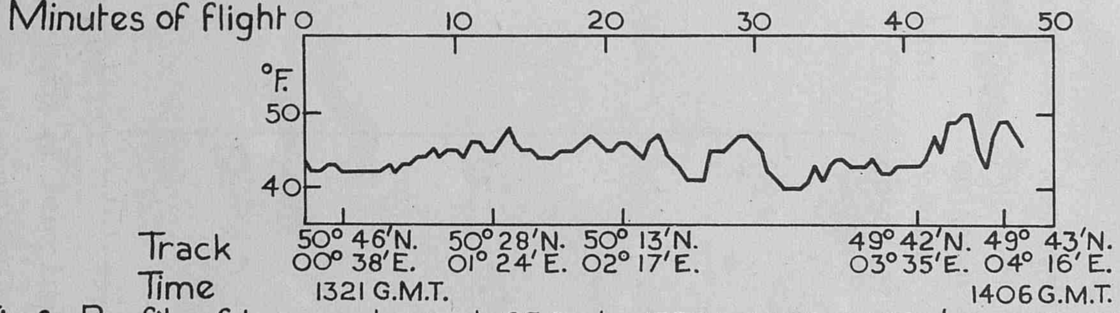

Fig.6 Profile of temperature at 850mb. 1320 to 1405 G.M.T. June 12th, 1951 on track from 50°46′N. 00°38′E. to 49°23′N. 04°16′E.
Example of large irregularities of temperature during a flight through a cold front

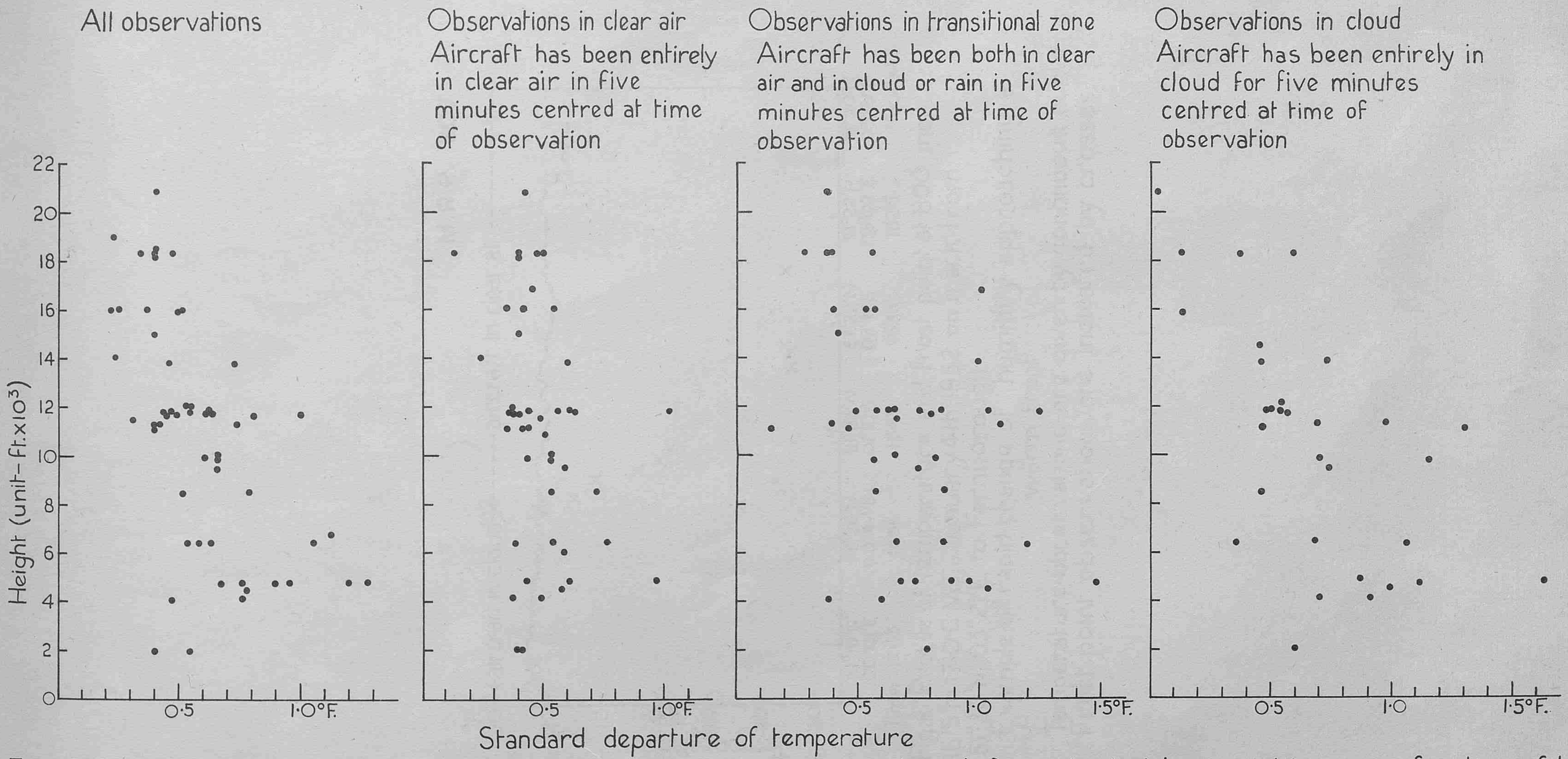

Fig.7 Standard departure of temperature from 5-minute running mean plotted for individual horizontal legs as a function of height

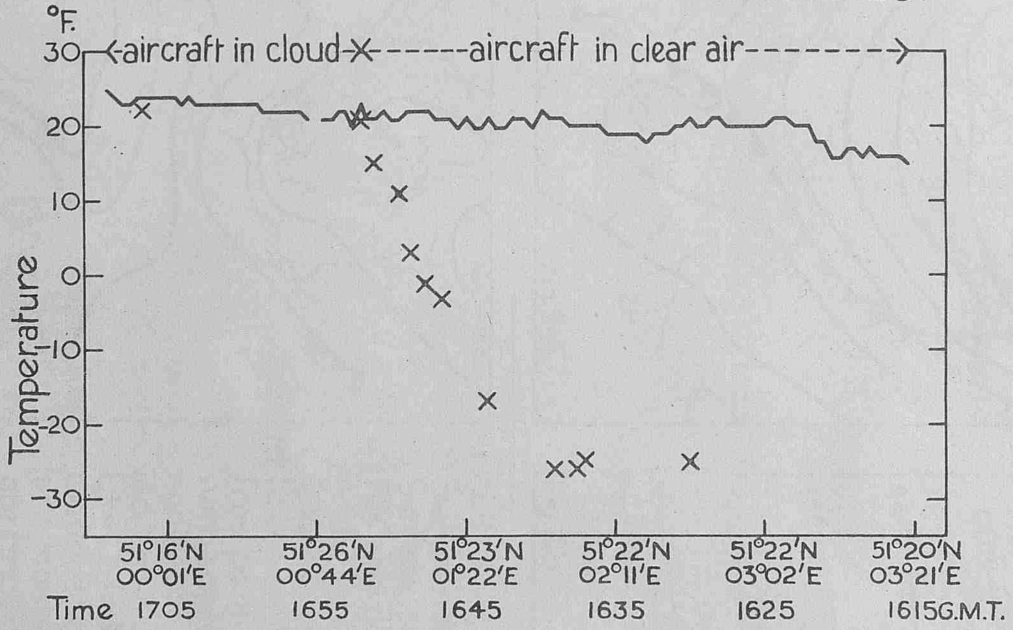

Fig.8 Profile of temperature and frost point at 800 mb. 1615 to 1710 G.M.T. January 4th, 1952 on track from 51°20'N. 03°21'E. to Farnborough

Example of rapid change of humidity approaching a warm front

Temperature observations are given by continuous curve
Frost point observations are indicated by crosses

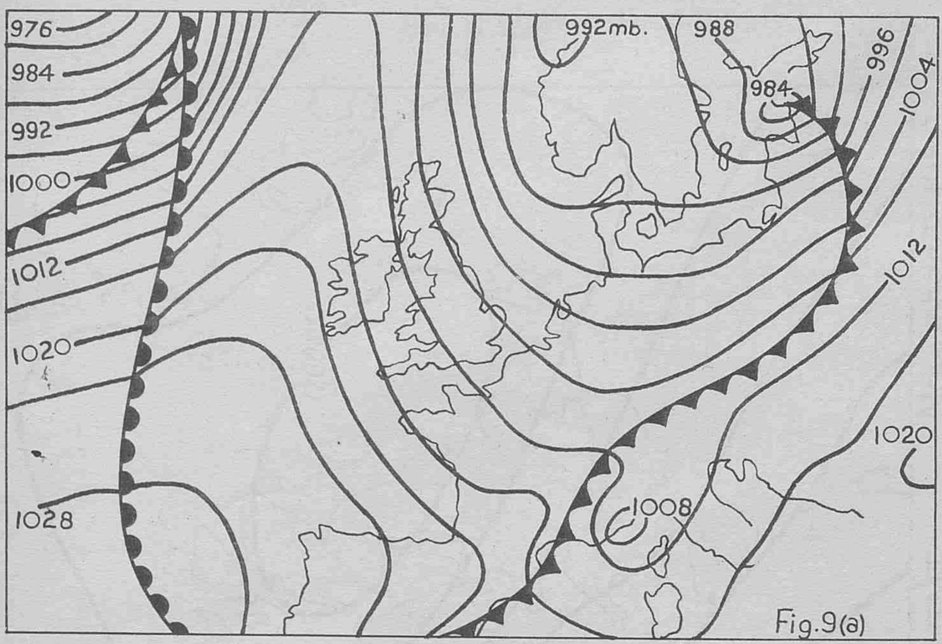
976
984
992
1000
1012
1020
1028
992mb.
988
984
996
1004
1012
1020
1008

Fig.9(a)

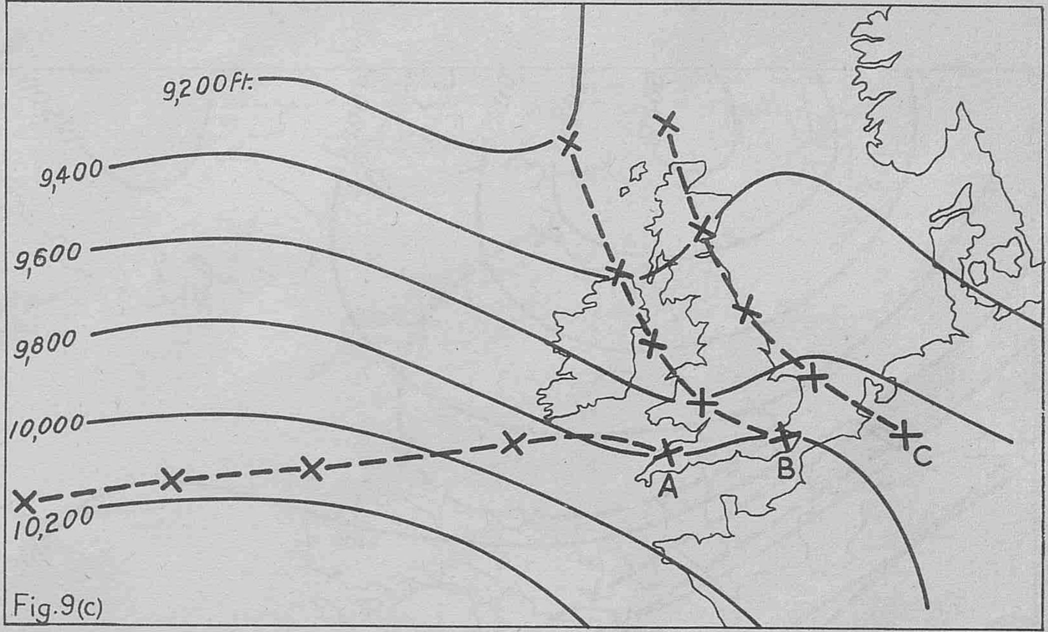
9,200 Ft.
9,400
9,600
9,800
10,000
10,200
A
B
C

Fig.9(c)

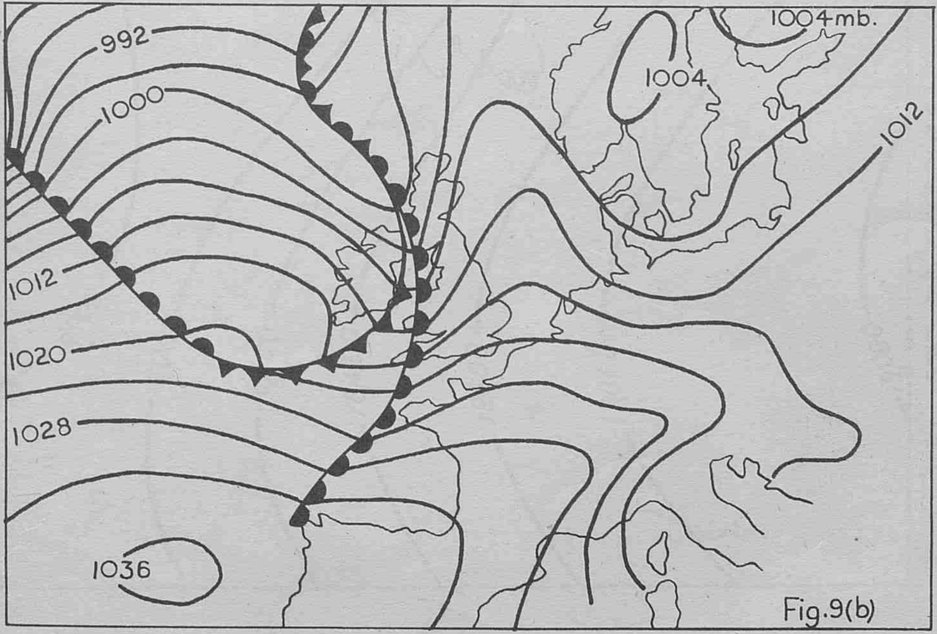
992
1000
1012
1020
1028
1036
1004
1004mb.
1012

Fig.9(b)

Fig.9

(a) Surface chart 1500 G.M.T. 3.1.52.
(b) Surface chart 1500 G.M.T. 4.1.52.
(c) 700 mb. contours 1500 G.M.T. 4.1.52 and 24 hr. trajectories at 700 mb. terminating at that time
(d) Cross-section of cloud and frostpoint at 1500 G.M.T. 4.1.52. with approximate positions of radio-sonde stations indicated (C=Camborne, Lk=Larkhill, B=Brussels). Boundaries of the frontal zone are indicated by heavy continuous lines.

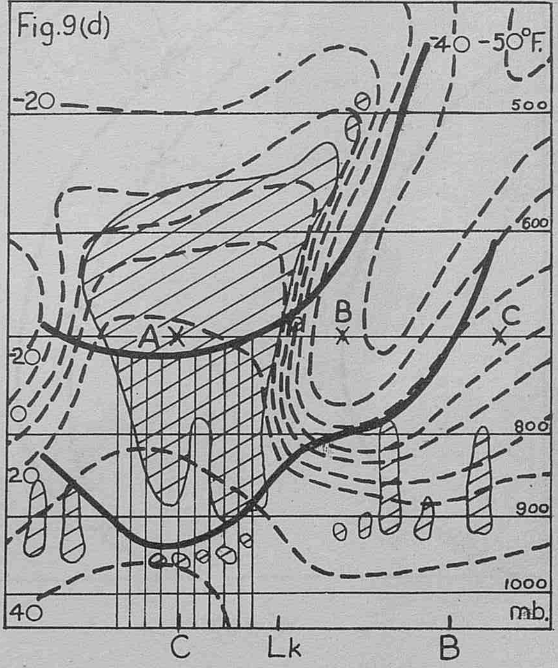

Fig.9(d)

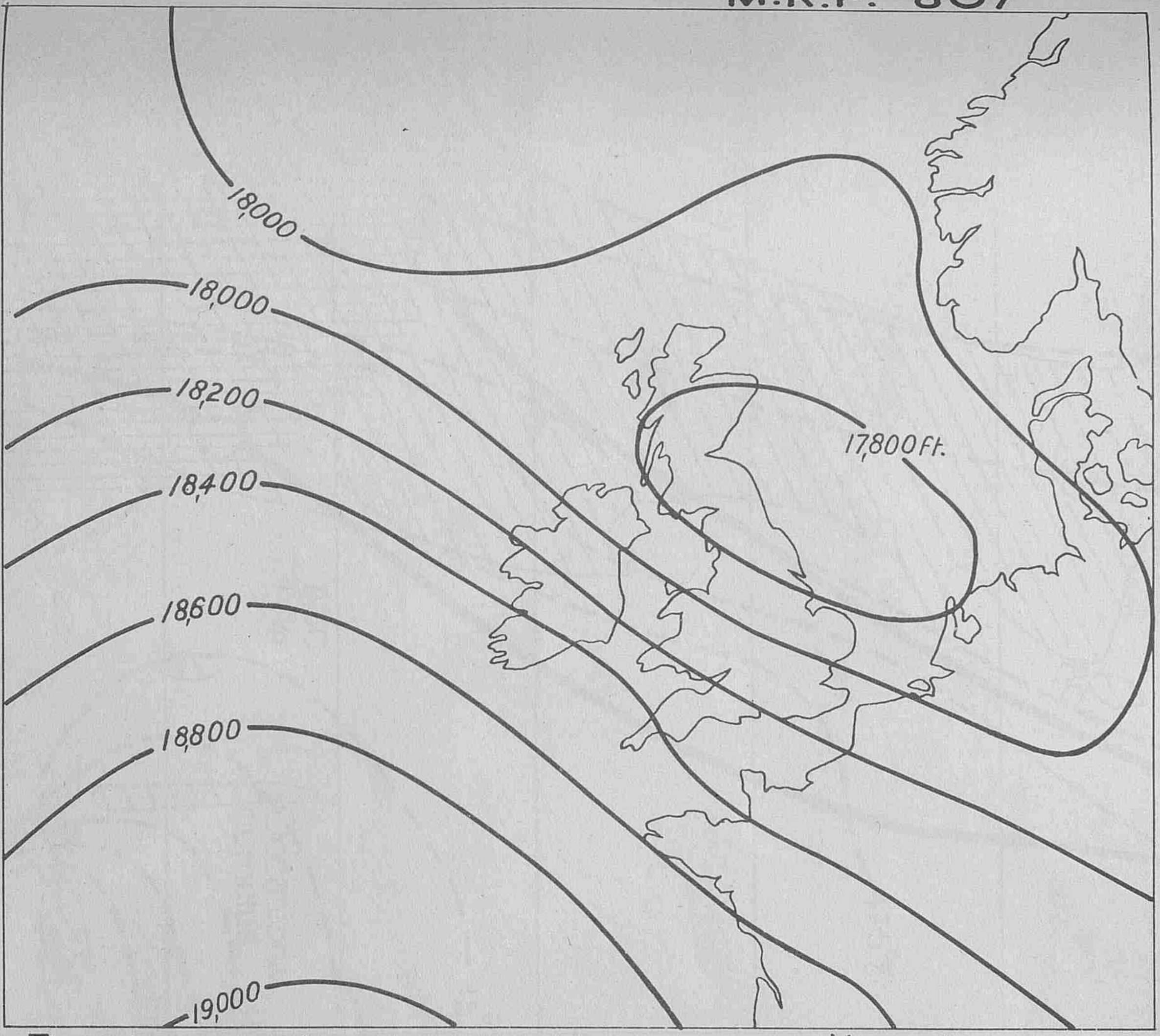
18,000
18,000
18,200
18,400
18,600
18,800
19,000
17,800 ft.

Fig.10(a) 500 mb. contours 1500 G.M.T. November 2, 1950

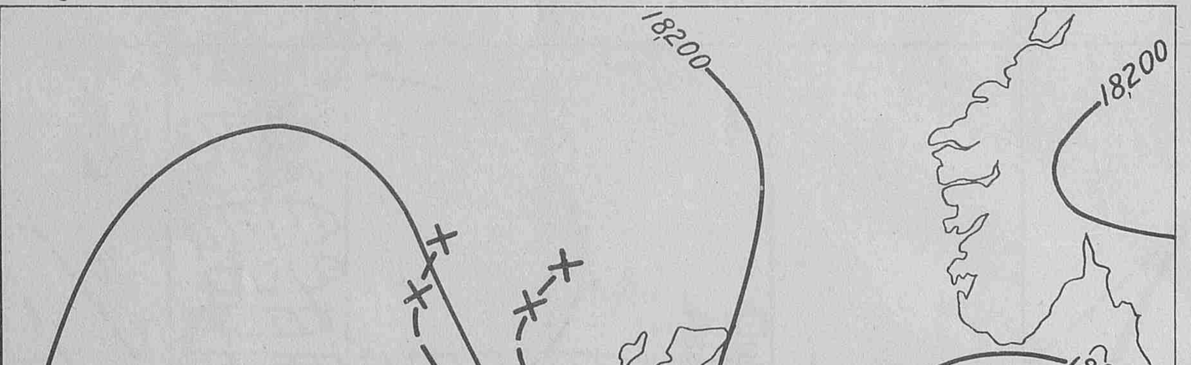
18,200
18,200
18,400
18,000
17,800
17,600
17,400
18,600
18,800 ft.

Fig.10(b) 500 mb. contours 1500 G.M.T. November 3, 1950
and 24-hour trajectories at 500 mb. terminating at that time
(crosses give position at 6-hour intervals)

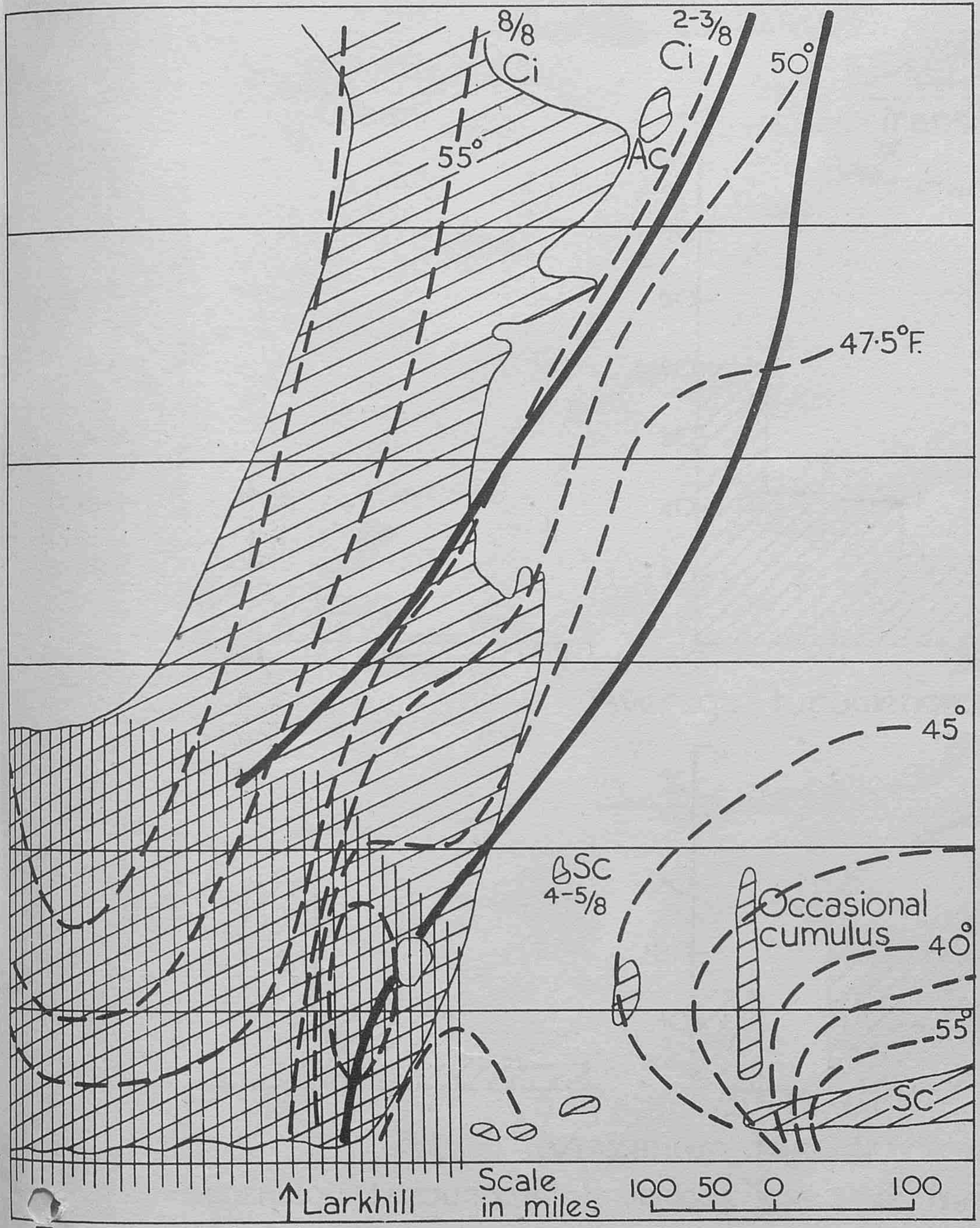
8/8
Ci
2-3/8
Ci
50°
55°
Ac
47·5°F.
45°
SC
4-5/8
Occasional
cumulus
40°
55°
Sc
↑Larkhill
Scale
in miles
100 50 0 100

Fig.11 March 6th, 1952 1500 G.M.T.

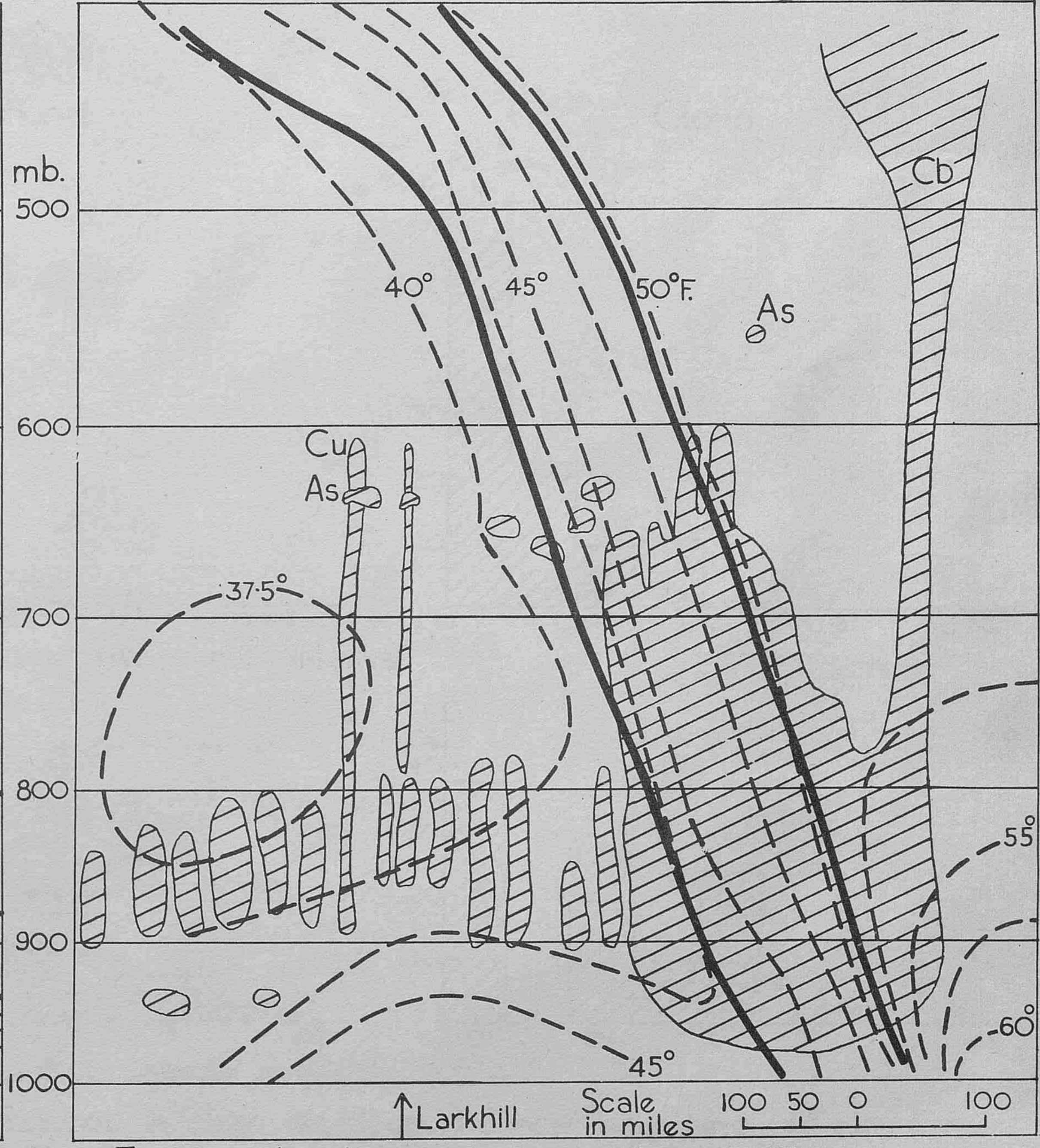
mb.
500
600
700
800
900
1000
Cb
40°
45°
50°F.
As
Cu
As
37·5°
55°
60°
45°
↑Larkhill
Scale
in miles
100 50 0 100

Fig.12 April 27th, 1951 1500 G.M.T.

Cross-section of cloud and wet bulb potential temperature required for saturation. Typical warm and cold fronts

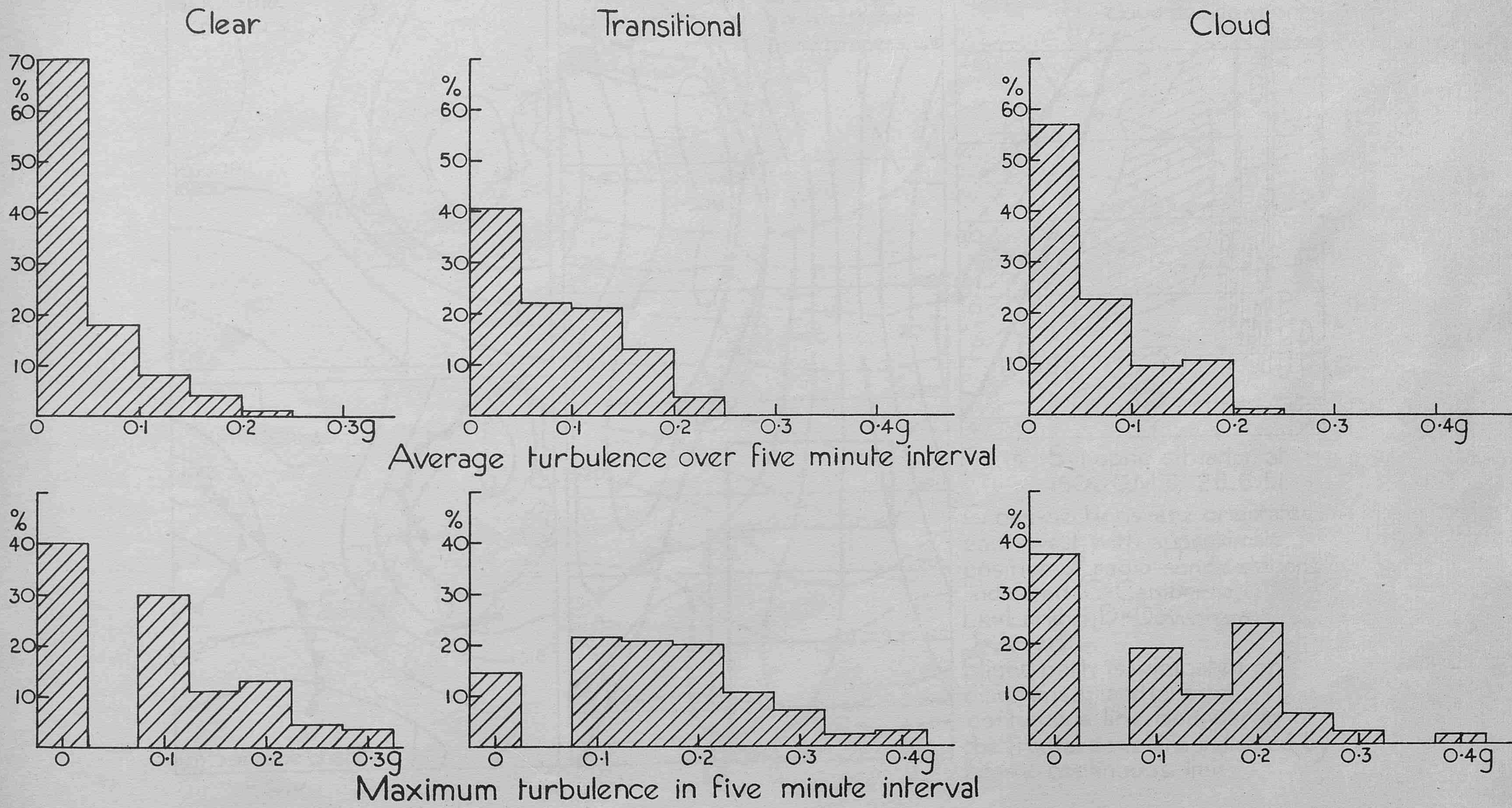

Fig.13 Percentage frequency of various levels of turbulence in clear air, cloud and near boundaries of cloud or precipitation (transitional)

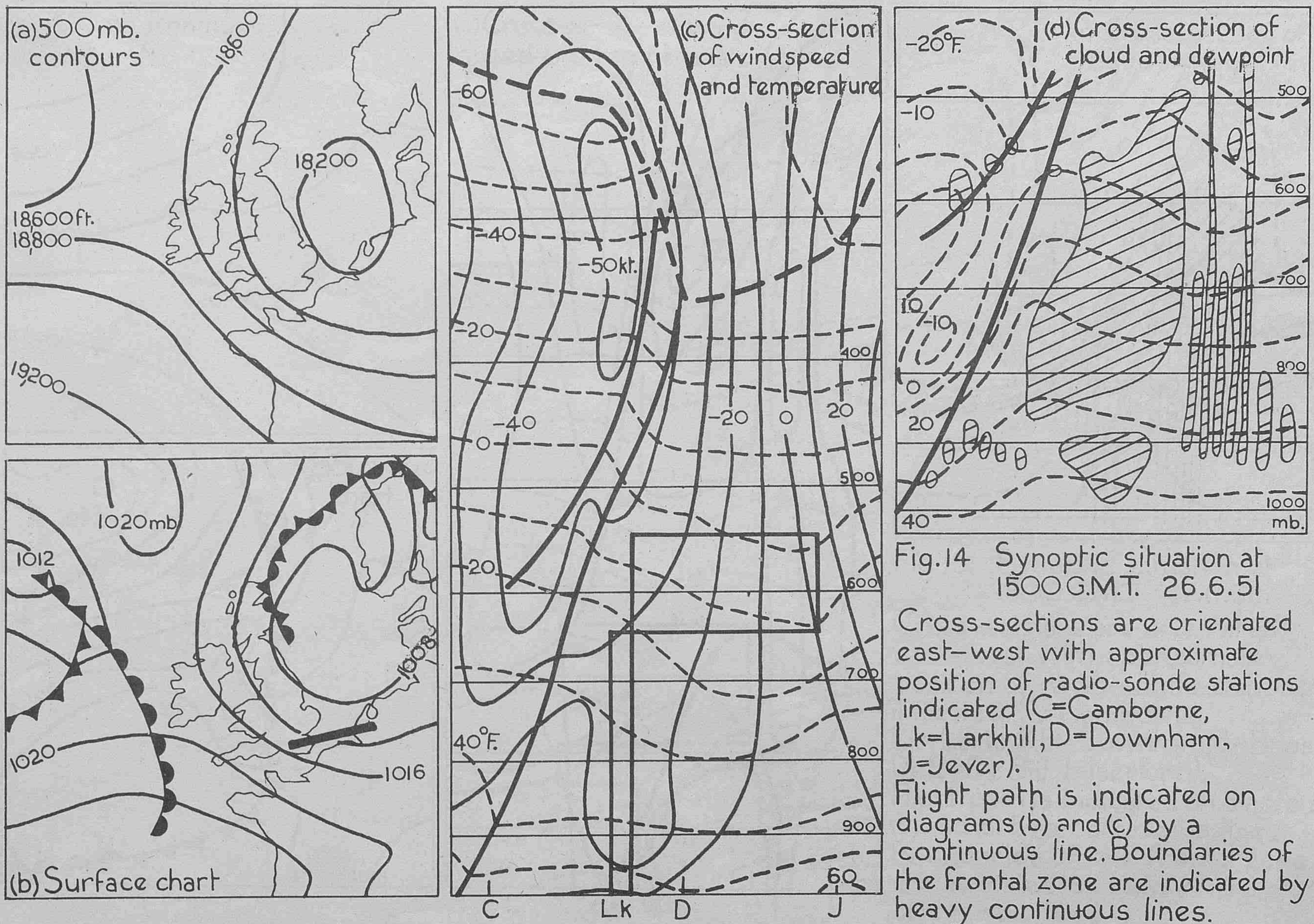
(a) 500 mb. contours
18600
18,200
18600 ft.
18,800
19,200
(b) Surface chart
1020 mb
1012
1020
1016
1008
(c) Cross-section of windspeed and temperature
-60
-40
-50 kt.
-20
-40
0
-20
-20
0
20
40°F.
400
500
600
700
800
900
60
C
Lk
D
J
(d) Cross-section of cloud and dewpoint
-20°F.
-10
10
-10
0
20
40
500
600
700
800
1000
mb.

Fig. 14 Synoptic situation at 1500 G.M.T. 26.6.51

Cross-sections are orientated east–west with approximate position of radio-sonde stations indicated (C=Camborne, Lk=Larkhill, D=Downham, J=Jever).

Flight path is indicated on diagrams (b) and (c) by a continuous line. Boundaries of the frontal zone are indicated by heavy continuous lines.

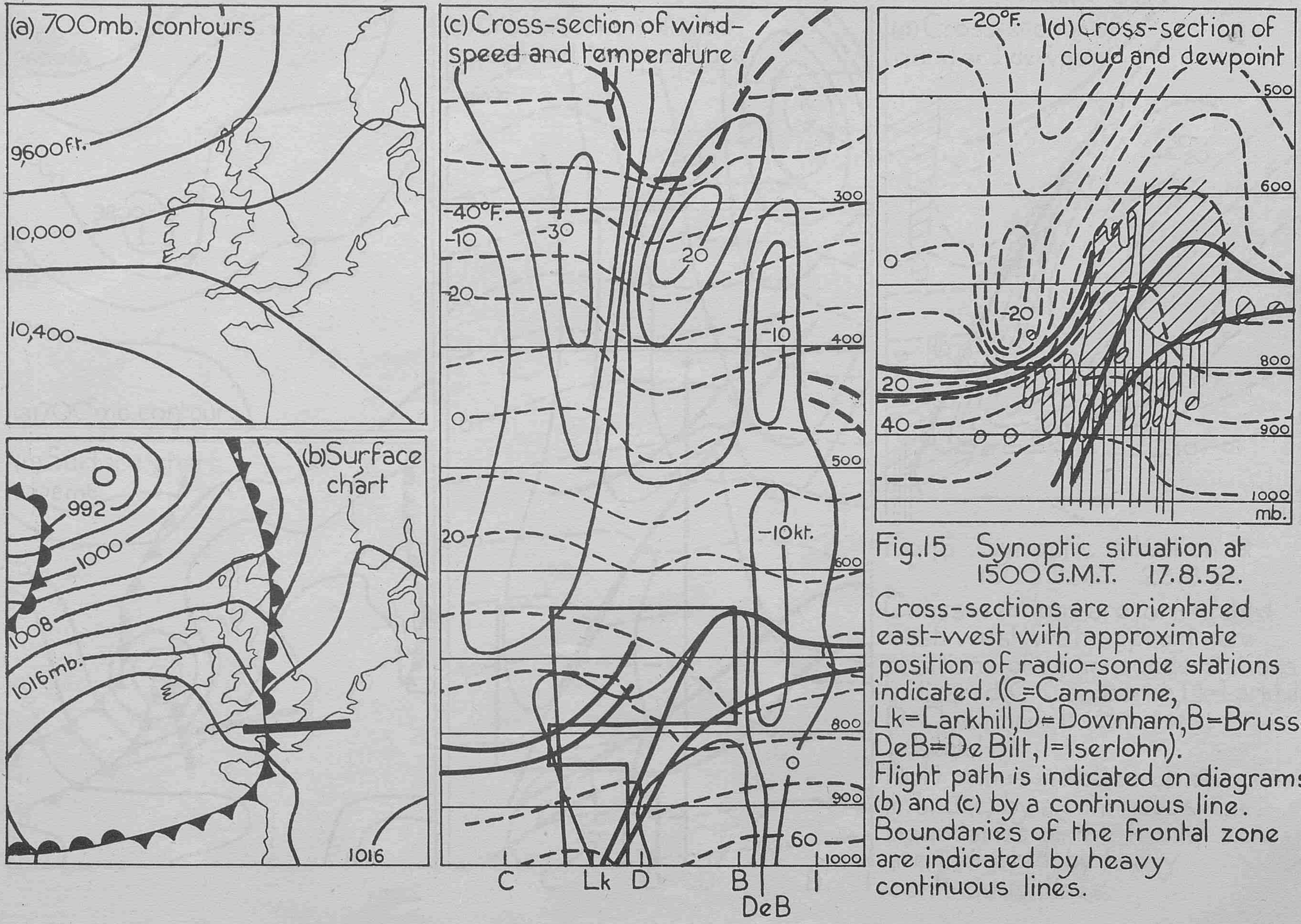
(a) 700mb. contours
9600ft.
10,000
10,400
(b) Surface chart
992
1000
1008
1016mb.
1016
(c) Cross-section of wind-speed and temperature
-40°F.
-10
-30
20
-20
-10
0
20
-10kt.
60
0
300
400
500
600
800
900
1000
C
Lk
D
B
DeB
I
-20°F.
(d) Cross-section of cloud and dewpoint
500
600
0
-20
800
20
40
900
1000
mb.

Fig.15 Synoptic situation at 1500 G.M.T. 17.8.52.

Cross-sections are orientated east-west with approximate position of radio-sonde stations indicated. (C=Camborne, Lk=Larkhill, D=Downham, B=Brussels, DeB=De Bilt, I=Iserlohn).
Flight path is indicated on diagrams (b) and (c) by a continuous line.
Boundaries of the frontal zone are indicated by heavy continuous lines.

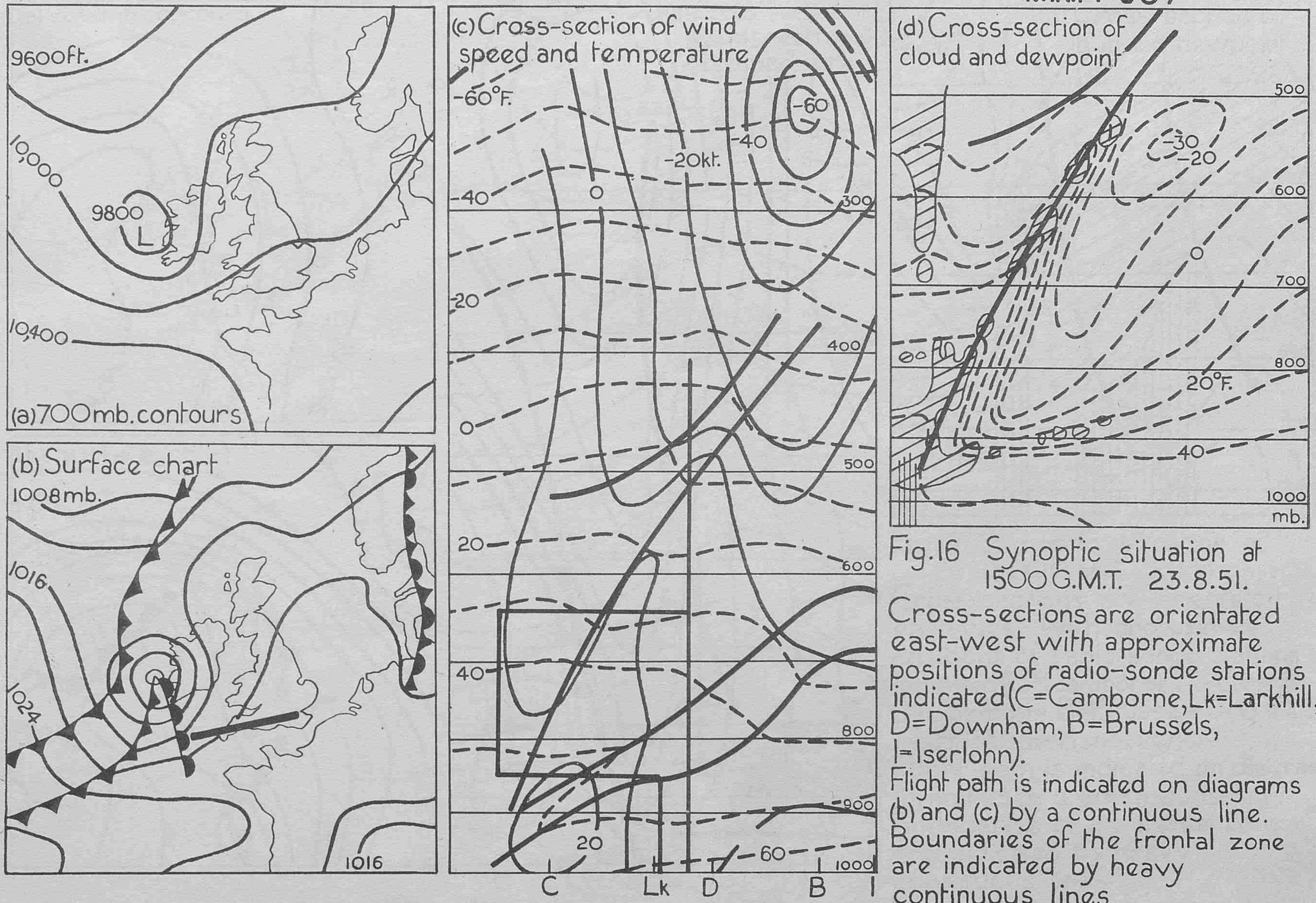

Fig.16 Synoptic situation at 1500 G.M.T. 23.8.51.

Cross-sections are orientated east-west with approximate positions of radio-sonde stations indicated (C=Camborne, Lk=Larkhill, D=Downham, B=Brussels, I=Iserlohn).

Flight path is indicated on diagrams (b) and (c) by a continuous line. Boundaries of the frontal zone are indicated by heavy continuous lines.

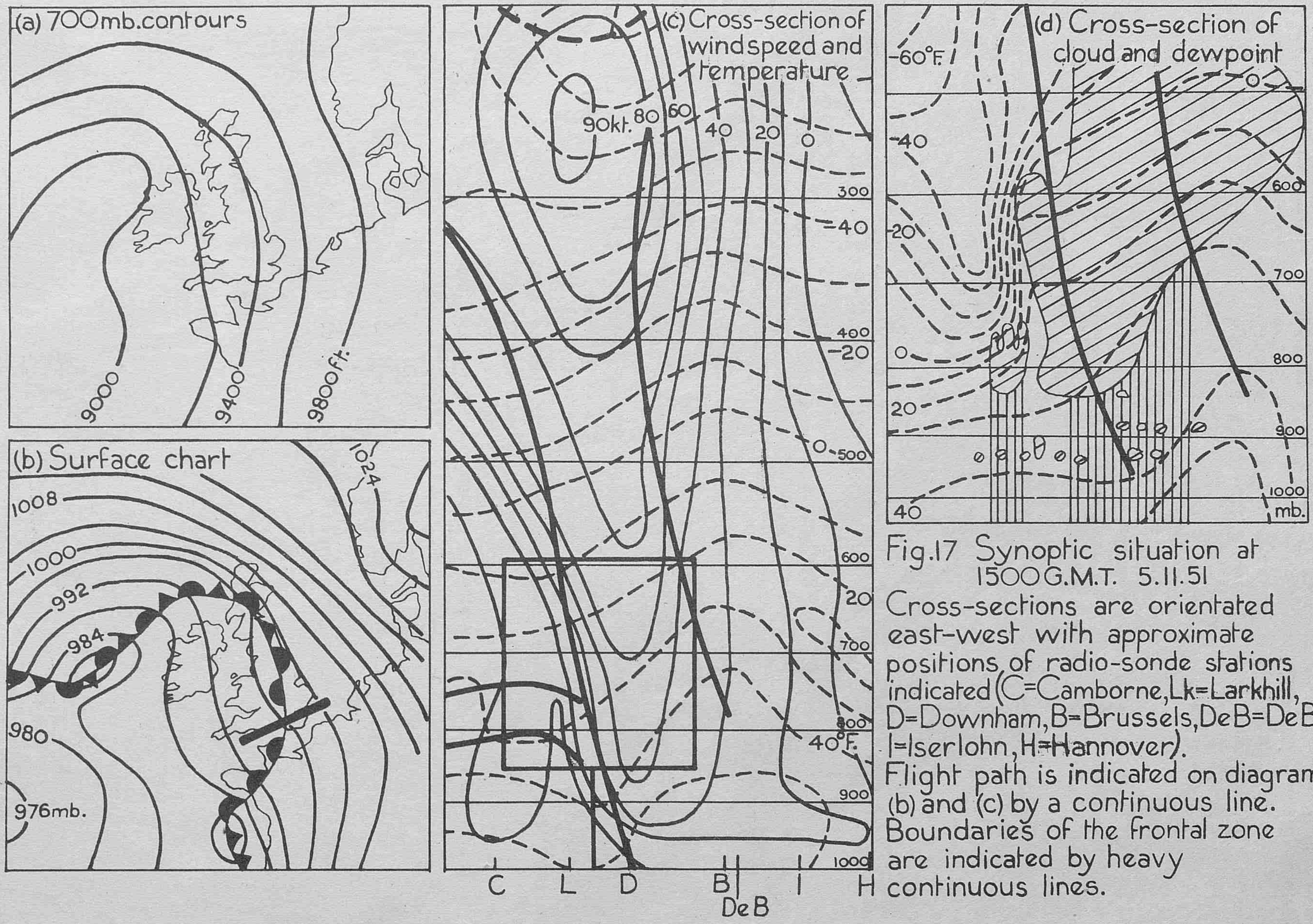

Fig.17 Synoptic situation at 1500 G.M.T. 5.11.51

Cross-sections are orientated east-west with approximate positions of radio-sonde stations indicated (C=Camborne, Lk=Larkhill, D=Downham, B=Brussels, DeB=De Bilt, I=Iserlohn, H=Hannover).
Flight path is indicated on diagrams (b) and (c) by a continuous line.
Boundaries of the frontal zone are indicated by heavy continuous lines.

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Committee (London),
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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

# Some examples of the energy spectrum of turbulence in the atmosphere near the ground.

by

G.D. Robinson (Meteorological Office)

## Summary.

A numerical frequency analysis has been made of three one minute records of various components of the air velocity at 150cm above a short grass surface; cases with large positive and negative temperature gradients being included. The analysis is considered adequate over a frequency range from about 1/5c. sec$^{-1}$ to about 8c. sec$^{-1}$. Over the range from about 3c. sec$^{-1}$ to 8c. sec$^{-1}$ the turbulence is found to be isotropic, and the energy to vary as the inverse square of frequency, in agreement with a hypothesis of Kolmogoroff.

The magnitude of certain terms in the energy equation of the turbulent motion is examined.

## 1. Introduction.

Measurements of the fluctuations of wind and temperature were begun at Kew Observatory in an attempt to measure the vertical convective flux of heat. Preliminary results (Lander and Robinson 1952) suggested that a numerically low value was being obtained systematically, and subsequent experience, not yet reported, confirms this. The measurements also allowed computation of the vertical convective flux of momentum (Reynolds stress), and the results of this computation were smaller than the expected stress, though there was no direct measurement for comparison. One possible explanation of these anomalies was that the size range of the eddies investigated was not wide enough to cover all those important in the transfer process, and in particular that eddies smaller than those detected might be important. It is planned to check this by measurement, but in the meantime it seemed worthwhile to analyse some of the records to determine the frequency spectra of the horizontal and vertical motions since there were indications that at the higher frequencies the motion might be approximately isotropic, and therefore not responsible for momentum transfer (Lander and Robinson 1952 p.14).

Such an analysis has a more general application. The hypotheses of Kolmogoroff (see e.g. Liepmann 1952) include the statement that the small scale motion is isotropic and, though there could be no question of "confirming" the hypothesis, or otherwise, it seemed that a demonstration of small scale isotropy

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M.R.P. 808
S.C.III/151
11th May, 1953

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TECHNICAL REFERENCE FILES

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

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Such an analysis has a more general application. The hypotheses of Kolmogoroff (see e.g. Liepmann 1952) include the statement that the small scale motion is isotropic and, though there could be no question of "confirming" the hypothesis, or otherwise, it seemed that a demonstration of small scale isotropy in the turbulence near the ground, should such a demonstration emerge, would have some interest. In particular it would give an indication of the nature and scale of those phenomena in the flow near the ground which might be investigated profitably by applying Kolmogoroff's hypotheses.

/No

No detailed publication of previous tests of the Kolmogoroff hypotheses in the atmosphere has been found. Priestley and Sheppard (1952) report work by R.J. Taylor, at Melbourne, and Liepmann (1952) mentions work by P.B. MacCready, at Pasadena, both of whom appear to have found isotropic turbulence on a scale up to about 100 cm, but details of their methods and of the situations in which they worked are not yet available.

## 2. Notation.

The following notation will be used.

| | |
|---|---|
| $x, u, i,$ | Displacement, instantaneous velocity component, and unit vector in the direction of the mean wind for the period examined. |
| $y, v, j,$ | The same for the direction perpendicular to the mean wind in the horizontal. |
| $z, w, k,$ | The same for the vertical direction. |
| $\rho, p, T,$ | Air density, pressure, and temperature at the point of observation. |
| $\bar{u}, \bar{v}, \bar{w}, \bar{\rho}, \bar{p}, \bar{T},$ | Arithmetic means of a number of successive observations at a point of the appropriate quantity. |
| $u'$ etc. | $u - \bar{u}$ etc. |
| $h, \bar{h}, h',$ | $\lvert iu + jv \rvert,$ etc, the magnitude of the horizontal vector component of the velocity, etc. |

Other symbols are defined in the text.

## 3. The measurements.

The description of the apparatus and procedure given by Lander and Robinson (1952) will not be repeated here. The material available for analysis comprised a number of sets of 200 readings, of $h$ and $kw$ spaced at intervals of 1/5 or 1/4 sec i.e. the longest period of continuous recording was 50 sec. It can readily be seen that the fluctuations of $h$ are comparable in magnitude with those of the dominant horizontal component, or with those of either horizontal component when these are approximately equal. It seemed advisable to confirm this and to increase the scope of the investigation by recordings of $u$ and $v$. Records were made on one occasion at a height of 130cm on the site previously used. The hot wire anemometer was clamped with the wires in the horizontal plane and the axis of symmetry along the estimated direction of the mean wind, a day with a fairly strong and steady wind being chosen. Recording and reduction of the traces followed the normal course, and the mean component across the anemometer during the run was computed. This allowed determination of the true direction of the mean wind, and of $u$ and $v$. I am indebted to Mr. A.J. Lander for his part in taking and reducing these records.

The lower limit of frequency which could be investigated was set by the length of the continuous record, and was estimated as about 1/5 c.sec$^{-1}$. The upper limit might have been set by the interval between measurements on the trace, 1/4 to 1/5 sec, but as will be seen later an apparently justifiable interpolation was practised, and the upper frequency limit was in fact determined by the time constants of the apparatus. It was confirmed experimentally (by Mr. Lander) that the time constant of the wire was a few milliseconds at the wind speeds investigated, and that the response of the apparatus was controlled, as planned, by the galvanometers, which had a free period of 1/10 sec and a damping constant adjusted to 0.7 critical to minimise the effects of distortion. Records had been taken at heights of 25cm and about 150cm, but it was clear from the preliminary examination that at 25cm a good deal of the energy of the vertical component was concentrated at frequencies too high to be well recorded, so only the 150cm measurements were considered. There was a further limitation on the choice of record. To specify the nature of the turbulence as completely as possible it was necessary to know the local rate of production of kinetic energy of turbulence, involving the terms

$$(g/c_pT)\ \overline{\rho}\ \overline{w'T'} \qquad \text{and} \qquad \overline{\rho}\ \overline{u'w'}\ \partial\overline{u}/\partial z$$

(Throughout this work the approximations $\overline{\rho}\ \overline{w'T'}$ and $\overline{\rho}\ \overline{u'w'}$ were perforce used for $\overline{\rho w'T'}$ and $\overline{\rho u'w'}$). The terms $\overline{\rho}\ \overline{w'T'}$ and $\overline{\rho}\ \overline{u'w'}$

had been computed for the 50 sec runs, and the gradient $\partial\overline{u}/\partial z$ measured over a 30 min. period during which several runs were made. The individual values, of $\overline{\rho}\ \overline{u'w'}$ in particular, varied considerably from run to run, and it was necessary to choose individual records for which the value of $\overline{\rho}\ \overline{u'w'}$ was approximately the same as the mean for the set, so that it could justifiably be associated with the long period mean of $\partial\overline{u}/\partial z$. There were other restrictions on the choice of record which will be mentioned in the following paragraph. Table I shows the details of the records finally chosen.

## 4. Reduction of the measurements. Results.

The method of analysis was the well known one, based on a theorem ("Wiener - Khinchin theorem") first applied to the turbulence problem by G.I. Taylor (1938). The material comprises a series of observations $a_1, a_2, .. a_m$ at equal intervals of time $\tau$. Correlation coefficients $R_a(p\tau) = \sum_m a_m a_{m+p} / \sum_m a_m^2$ are formed for p = 1, 2, etc. and the function $R_a(t)$ formed by interpolation. If now $F_a(n)$ is the normalised energy spectrum function, i.e. if $\overline{a_m^2} F_a(n)\,dn$ denotes the square of the amplitude of the component of a contained between frequencies n and n+dn , the theorem states

$$F_a(n) = 4\int_0^\infty R_a(t)\cos 2\pi nt\,dt\ .$$

Two conditions of importance in the present application are implied; firstly that the series of a is stationary, or homogeneous in time, so that the values of the correlation coefficients do not depend on the point in the series from which they are computed, and secondly that $\int_0^\infty R_a(t)dt$ reaches a limiting value in a time small compared with the extent in time of the series of a .

It is doubtful how far these conditions are ever satisfied when the a's are measurements made in the atmosphere. They were clearly not satisfied in many of the runs of 200 observations available for analysis, but a number of the runs appeared to contain a fairly homogeneous set of fluctuations about a fairly steady mean. The considerations detailed in para. 3 had already greatly restricted the choice, and the two occasions chosen for analysis of w and h were not ideal, as will be seen. There was considerable long period variation present in the record chosen for analysis of u and v , though it appeared to be the most homogeneous of the 10 records available.

Ignoring these departures from the ideal the correlation coefficients were computed for the six series chosen, for each time interval or at least for alternate intervals up to about the 12th (i.e. t = 3 sec about) and for two or three longer intervals. More than the 200 observations would have been required to give reliable detail at longer intervals, quite apart from the labour involved (all computation was done on a hand operated keyboard machine). Smooth curves were drawn through the points obtained, ordinates of the R(t) curve read off, and the Fourier transform computed numerically from n = 8 to n = 1/4, and roughly estimated at two or three points between n = 1/4 and n = 0. Interpolation of the R(t) curve had, of course, a controlling effect on the results for large n , so three independent curves of R(t) were drawn to different scales, and interpolation at intervals of 1/64 sec, 1/60 sec, and 1/40 sec made on the different curves. The F (n) curves are thus composed of points computed from three independent interpolations, and on each F (n) curve there is a point which has been computed from more than one interpolation. It is of course assumed that there is no fine detail on the R(t) curves on a time scale less than 1/4 sec. The only justification which can be advanced for this (other than its apparent reasonableness) is that it was confirmed in a single instance when a run (at a height of 25cm) was measured at 1/10 sec intervals and the first two or three correlations computed. An overall check on the computation was available, since we have

$$\int_0^\infty dn \int_0^\infty R(t) \cos 2\pi nt \, dt = \frac{1}{4}\int_0^\infty F(n)\, dn \equiv \frac{1}{4} .$$

This check was applied, except in the analysis of the two horizontal components, when reliable values of F (n) could not be obtained for small n . The computed correlation coefficients are given in Table II, and the spectrum functions in Table III and Figs. 1, 2 and 3. The values of F (n) for n = 8 are less well determined than for other frequencies, the number of ordinates used in computation of the Fourier integral being smaller. They are comparable amongst themselves, but not with values for the other frequencies. The correction term shown in Table III was applied to adjust the intensities for distortion in recording. They are theoretical factors appropriate to a system of free period 1/10 sec and damping constant 0.7 critical.

Figs. 1, 2 and 3 show that in each case there is at the higher frequencies an approximate equipartition of energy between the turbulent components investigated. The low frequency limit of the isotropy appears to be about 3 to 4 $c.sec^{-1}$, and the two horizontal components appear to be of approximately equal energy at frequencies higher than about 1.5 $c.sec^{-1}$ in the one not very satisfactory case investigated. The results of the analysis of the horizontal components are regarded as establishing approximately the equivalence of the analysis of h' and u' (since the total energy and the first correlation coefficient are approximately equal) and the equipartition of energy between u' and v' at the higher frequencies, but it must be remembered that the series of u' was not really suitable for analysis.

## 5. Kolmogoroff's hypotheses.

These are stated by Liepmann (1952) as follows:-

(1) The small scale motion is always isotropic.

(2) The energy spectrum of the small scale motion can not depend upon details of the flow ---(it) will depend upon the dissipated energy per unit volume and time, $\epsilon$ , and the viscosity, $\nu$ .

(3) If the Reynolds number of the problem is sufficiently large the zone of dissipation and the zone of production of turbulent energy will be widely separated in the wave number space. Then there may exist a range of wave numbers which are in a state of local isotropic equilibrium but which are still not in the dissipation region. In this (region the energy spectrum) will become independent of $\nu$ .

We have seen that the present results are consistent with the first hypothesis and now proceed to examine their relation to the second and third. It is to be noted in the first place that our analysis has been made in terms of frequency, differing in this respect from most laboratory analyses of turbulent flow in a wind tunnel. The correlations which we have computed are the correlations of the velocities at a fixed point separated by a given time, averaged in time. The correlations normally obtained in wind tunnel work (those most readily obtained automatically) are between velocities at points separated by a given distance, averaged in time. Our observations allow analysis in terms of frequency only; the more usual observations, on the assumption that the average in time is equivalent to an average over a large number of similarly situated pairs of points in space, allow analysis in terms of wave number.

The form of the energy spectrum in the region covered by the third Kolmogoroff hypothesis, if this region exists, is readily obtained by dimensional arguments. In terms of frequency it is $F(n) = C\epsilon n^{-2}$ and in terms of wave number k $F(k) = C\,\epsilon^{2/3}\,k^{-5/3}$.

(We note that the hypothesis is not consistent with the relation $n = ck$ where $c$ is a constant "phase velocity", though this assumption has sometimes been made with $c$ equated to the mean flow velocity). The observations in the region of approximate isotropy were therefore examined for a power law relation between $F(n)$ and $n$, by taking logarithms and finding the best fitting linear relation by least squares. The results are shown in Table IV. The inverse square relation is indicated without ambiguity.

These results suggest that the frequency region between about 3c. $\mathrm{sec}^{-1}$ and the high frequency limit at 6-8c. $\mathrm{sec}^{-1}$ has the properties postulated in the third Kolmogoroff hypothesis. It would be desirable to check that viscous dissipation in this region was negligible in the cases considered, and that $F(n)$ was directly proportional to the total rate of viscous dissipation $\epsilon$. Unfortunately an analysis in frequency gives no information on the dissipation, which involves terms containing $\partial w/\partial x$ etc, and necessitates an analysis in space. Without some relation between frequency and wave number the present results cannot be used to determine the contribution to $\epsilon$ of the frequencies investigated, and, as we have seen, the linear relation sometimes assumed is not consistent with the Kolmogoroff hypothesis.

## 6. The energy balance of the turbulent motion.

Determination of the total viscous dissipation involves consideration of the energy balance of the turbulent motion. Calder (1949) has given a convenient approximate analysis in which are made assumptions regarding the fluctuations of density essentially similar to those adopted in the present investigation. He gives the energy balance of the turbulent motion (his eqn.35) in the following form:-

Local rate of increase of turbulent kinetic energy

= Transfer of energy from mean motion (terms like $\rho\,\overline{u'w'}\,\partial\overline{u}/\partial z$)

+ Advection of turbulent energy by the mean motion

- Viscous dissipation $\epsilon$

+ Work done by gravity consequent on heat transport by turbulence $((g\overline{\rho}/\overline{T})\,\overline{w'T'})$

+ Work done by the fluctuating gradients of pressure

+ Transport (diffusion) of turbulent energy by the turbulent motion (terms like $\frac{1}{2}\,\overline{\rho}\,\partial(\overline{w'u'^2})/\partial z$ ).

We may assume the local change of turbulent energy to be zero, though even if there is no steady increase or decrease large fluctuations clearly occur. We may also assume no advection, with similar lack of precision. We have assumed from the start that the pressure fluctuations are negligible, but the assumption is quite unwarranted in the present application. The "pressure-fluctuation term" in the energy balance has a form very similar to the "diffusion" term and may be of the same magnitude.

The terms $\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z$ and $(g\bar{\rho}/\bar{T})\ \overline{w'T'}$ are available in Table I, but it must be remembered that the heat flux implied by these figures is almost certainly about half the true value, and the momentum flux is probably also too small.

The term $\bar{\rho}\ (\overline{w'h'^2}+\overline{w'^3})/2$ was computed for the cases examined; its divergence cannot be derived so an idea of its order of magnitude was obtained by assuming a linear change with height from a zero at the surface. Table V shows the terms in the energy balance so far as they can be determined [*]. It is clearly out of the question to attempt to derive the viscous dissipation by difference, though its order of magnitude is indicated.

It was thus not possible to check the relation between F(n) and $\epsilon$, but one curious circumstance seems worthy of record. The ratio of the constants in the equation $F(n) = An^{-2}$ for the occasions for which the vertical velocities were available is, from Table IV, 0.35/0.115 = 3.04. The ratio of the terms $\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z$ is, from Table V, 2.4/0.78 = 3.08. This fact is further illustrated in Fig. 4, where individual values of $F(n)/(\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z)$ are plotted against n and in relation to the curve $F(n) = 20\ n^{-2}$. If this is a coincidence it is a rather remarkable one. It is tempting to advance the hypothesis----
"When the turbulent energy does not change with time energy supplied through the terms $\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z$ etc. is dissipated locally, energy supplied or removed by the heat flux term $(g\bar{\rho}/\bar{T})\ \overline{w'T'}$ is removed or supplied by advection or diffusion". This is very different from the "Richardson" picture and of course far more evidence, including a clarification of the meaning of $\bar{\rho}\ \overline{w'T'}$ and $\bar{\rho}\ \overline{h'w'}$ as determined by the present method, is required before it can be considered seriously, but it is not physically unreasonable.

## 7. Conclusions.

We may summarize the results of this investigation as follows.

If the turbulent velocity fluctuations are defined as deviations from a 1 minute mean and are analysed with respect to frequency, then at about 150cm above a short grass surface there is equipartition of energy between the three components of turbulent motion at frequencies above about 3c. sec$^{-1}$ at least to the limit of observation at 8c. sec$^{-1}$. The two horizontal components may be of equal energy at frequencies above about 1.5c. sec$^{-1}$. Between frequencies of 3c. sec$^{-1}$ and 6c. sec$^{-1}$ for the vertical component and 1.5c. sec$^{-1}$ and 6c. sec$^{-1}$ for the horizontal components (the upper limit being set by the method of analysis) there is a clearly defined relation of the form $F(n) \propto n^{-2}$. This is in accord with a hypothesis of Kolmogoroff and suggests that the motion at these frequencies is responsible for negligible viscous dissipation and is receiving energy only by transfer from longer-scale turbulent motion i.e. is not a carrier of the heat or momentum flux.

There is also a suggestion that the local rate of viscous dissipation may balance the local rate of working of the Reynolds stress, as determined in the approximate form $\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z$ by the method of observation adopted, in spite of the existence of a comparable rate of working of the buoyancy forces. This is not well established, and should be regarded as no more than a hint for future investigation.

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[*] It is interesting to note the great difference between these cases and those reported by R.J. Taylor (1952). The present instances are for light winds and extreme temperature gradients; the "Richardson number" on 11 Nov. is 1. Taylor's examples were for much more nearby adiabatic conditions.

On the point of immediate interest these conclusions suggest that there is little likelihood that the anomalous values for the heat flux which the present method yields can be due to the effects of motions too rapid to be detected by the instrument. Apparatus for a more direct test has however been assembled and will be used in the near future.

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Taylor, G.I. 1938 P.R.S. A164, p. 476.

Taylor, R.J. 1952 Q.J.R. Met. Soc., 78. p. 179.

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| | | | |
|---|---|---|---|
| $\overline{w}$ | + 6.15 cm sec$^{-1}$ | $\overline{h'^2}$ | 531 cm$^2$ sec$^{-2}$ |
| $\overline{h'w'}$ | - 34.3 cm$^2$ sec$^{-2}$ | $\overline{w'^2}$ | 233 cm$^2$ sec$^{-2}$ |

Mean values for the period 1645 to 1715 G.M.T.

| | | |
|---|---|---|
| $\overline{h'w'}$ | - 33 cm$^2$ sec$^{-2}$ | (6 runs) |
| $\overline{w'T'}$ | - 0.8 °C cm sec$^{-1}$ | (6 runs) |
| $\partial\overline{u}/\partial z$ | + 0.19 sec$^{-1}$ | |
| $\partial\overline{T}/\partial z$ | + 5.0x10$^{-3}$ °C cm$^{-1}$ | |

3. Date and Time. 27 January, 1953. ca. 1430 G.M.T.
Elements analysed. u, v and h.
Height above ground. 130 cm.
Mean values for record examined

| | |
|---|---|
| $\overline{u}$ | 442 cm sec$^{-1}$ |
| $\overline{u'^2}$ | 8635 cm$^2$ sec$^{-2}$ |
| $\overline{h'^2}$ | 8535 cm$^2$ sec$^{-2}$ |
| $\overline{v'^2}$ | 4680 cm$^2$ sec$^{-2}$ |

Long term mean values

| | |
|---|---|
| $\partial\overline{u}/\partial z$ | not determined |
| $\partial\overline{T}/\partial z$ | approximately adiabatic. |

n and in relation to the curve $F(n) = 20\ n^{-2}$ . If this is a coincidence it is a rather remarkable one. It is tempting to advance the hypothesis---- "When the turbulent energy does not change with time energy supplied through the terms $\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z$ etc. is dissipated locally, energy supplied or removed by the heat flux term $(g\bar{\rho}/\bar{T})\ \overline{w'T'}$ is removed or supplied by advection or diffusion". This is very different from the "Richardson" picture and of course far more evidence, including a clarification of the meaning of $\bar{\rho}\ \overline{w'T'}$ and $\bar{\rho}\ \overline{h'w'}$ as determined by the present method, is required before it can be considered seriously, but it is not physically unreasonable.

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There is also a suggestion that the local rate of viscous dissipation may balance the local rate of working of the Reynolds stress, as determined in the approximate form $\bar{\rho}\ \overline{h'w'}\ \partial\bar{h}/\partial z$ by the method of observation adopted, in spite of the existence of a comparable rate of working of the buoyancy forces. This is not well established, and should be regarded as no more than a hint for future investigation.

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# Table I.

Records examined.

1. Date and Time. 22 July, 1952. ca. 1420 G.M.T.
   Elements analysed. w and h.
   Height above ground. 135 cm.
   Mean values for record examined

| | | | |
|---|---|---|---|
| $\overline{h}$ | 214 cm sec$^{-1}$ | $\overline{w'T'}$ | + 10.6 °C cm sec$^{-1}$ |
| $\overline{w}$ | + 2.93 cm sec$^{-1}$ | $\overline{h'^2}$ | 4526 cm$^2$ sec$^{-2}$ |
| $\overline{h'w'}$ | - 116 cm$^2$ sec$^{-2}$ | $\overline{w'^2}$ | 893 cm$^2$ sec$^{-2}$ |

Mean values for the period 1412 to 1452 G.M.T.

| | | |
|---|---|---|
| $\overline{h'w'}$ | - 103 cm$^2$ sec$^{-2}$ | (6 runs) |
| $\overline{w'T'}$ | + 9.9 °C cm sec$^{-1}$ | (6 runs) |
| $\partial\overline{u}/\partial z$ | + 0.17 sec$^{-1}$ | |
| $\partial\overline{T}/\partial z$ | - 3.3x10$^{-3}$ °C cm$^{-1}$ | |

2. Date and Time. 11 November, 1952. ca. 1705 G.M.T.
   Elements analysed. w and h.
   Height above ground. 143 cm.
   Mean values for record examined

| | | | |
|---|---|---|---|
| $\overline{h}$ | 121 cm sec$^{-1}$ | $\overline{w'T'}$ | - 1.7 °C cm sec$^{-1}$ |
| $\overline{w}$ | + 6.15 cm sec$^{-1}$ | $\overline{h'^2}$ | 531 cm$^2$ sec$^{-2}$ |
| $\overline{h'w'}$ | - 34.3 cm$^2$ sec$^{-2}$ | $\overline{w'^2}$ | 233 cm$^2$ sec$^{-2}$ |

Mean values for the period 1645 to 1715 G.M.T.

| | | |
|---|---|---|
| $\overline{h'w'}$ | - 33 cm$^2$ sec$^{-2}$ | (6 runs) |
| $\overline{w'T'}$ | - 0.8 °C cm sec$^{-1}$ | (6 runs) |
| $\partial\overline{u}/\partial z$ | + 0.19 sec$^{-1}$ | |
| $\partial\overline{T}/\partial z$ | + 5.0x10$^{-3}$ °C cm$^{-1}$ | |

3. Date and Time. 27 January, 1953. ca. 1430 G.M.T.
   Elements analysed. u, v and h.
   Height above ground. 130 cm.
   Mean values for record examined

| | |
|---|---|
| $\overline{u}$ | 442 cm sec$^{-1}$ |
| $\overline{u'^2}$ | 8635 cm$^2$ sec$^{-2}$ |
| $\overline{h'^2}$ | 8535 cm$^2$ sec$^{-2}$ |
| $\overline{v'^2}$ | 4680 cm$^2$ sec$^{-2}$ |

Long term mean values

| | |
|---|---|
| $\partial\overline{u}/\partial z$ | not determined |
| $\partial\overline{T}/\partial z$ | approximately adiabatic. |

Table II. Correlation coefficients $R_a(p\tau)$ for various elements a and intervals $p\tau$. $\tau = 1/4$ sec.

| Date | Element | p |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
|  |  | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 20 | 25 | (50) |
| 22 Jul. 52 | h' | .890 | .828 |  | .740 |  | .647 |  | .542 |  | .422 |  |  |  |  | .257 | .04 |  |  |
|  | w' | .543 | .341 | .264 | .128 | .109 | .090 | .024 | -.039 | -.006 | -.007 |  |  |  |  |  |  |  |  |
| 11 Nov. 52 | h' | .716 | .582 | .461 | .356 | .294 | .251 |  | .246 |  | .208 |  | .195 |  |  | .179 | .180 | .202 | (-.10) |
|  | w' | .537 | .421 | .342 | .281 | .286 | .153 | .207 | .176 | .116 | .081 | .011 | -.002 | -.063 | -.106 | -.117 | -.064 | -.050 |  |
| 27 Jan. 53 | h' | .805 |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
|  | u' | .835 | .697 |  | .529 |  | .529 |  | .460 |  | .399 |  | .311 |  |  | .153 |  | .228 |  |
|  | v' | .616 | .406 |  | .256 |  | .256 |  | .179 |  | .121 |  | .079 |  |  | .083 |  | -.055 |  |

Table III. Energy density $\rho\,\overline{a^2}\,F_a(n)$ erg cm$^{-3}$/ c.sec$^{-1}$

| n | Correction Applied | a | | | | | |
|---|---|---|---|---|---|---|---|
| | | 22 Jul. 52 | | 11 Nov. 52 | | 27 Jan. 53 | |
| c.sec$^{-1}$ | | h' | w' | h' | w' | u' | v' |
| | 1.0 | (46.4) | (2.03) | (4.40) | (0.64) | | |
| 1/40 | 1.0 | | | (5.30) | (0.65) | | |
| 1/24 | 1.0 | (42.8) | | | | | |
| 1/16 | 1.0 | | | (1.61) | (0.93) | | |
| 1/12 | 1.0 | (22.5) | | | | | |
| 1/8 | 1.0 | 9.4 | 1.88 | 0.63 | 0.78 | | |
| 1/4 | 1.0 | 2.17 | 1.26 | 0.36 | 0.11 | | |
| 1/2 | 1.0 | 0.67 | 0.46 | 0.29 | 0.11 | | |
| 3/4 | 1.0 | | | 0.18 | $5.52\times10^{-2}$ | | |
| 1 | 1.01 | 0.255 | 0.131 | $7.87\times10^{-2}$ | $7.84\times10^{-2}$ | 6.38 | 8.51 |
| 3/2 | 1.02 | | 0.170 | $4.98\times10^{-2}$ | $3.15\times10^{-2}$ | 5.64 | 5.44 |
| 2 | 1.02 | $7.43\times10^{-2}$ | $2.84\times10^{-2}$ | $2.29\times10^{-2}$ | $1.28\times10^{-2}$ | 3.03 | 2.94 |
| 5/2 | 1.03 | $6.72\times10^{-2}$ | $3.53\times10^{-2}$ | $1.85\times10^{-2}$ | $2.06\times10^{-2}$ | 0.98 | 1.68 |
| 3 | 1.04 | $3.67\times10^{-2}$ | $3.35\times10^{-2}$ | $1.25\times10^{-2}$ | $7.13\times10^{-3}$ | 0.96 | 1.31 |
| 4 | 1.08 | $1.62\times10^{-2}$ | $2.30\times10^{-2}$ | $7.96\times10^{-3}$ | $5.64\times10^{-3}$ | 0.71 | 0.97 |
| 5 | 1.13 | $1.71\times10^{-2}$ | $1.15\times10^{-2}$ | $3.68\times10^{-3}$ | $4.43\times10^{-3}$ | 0.25 | 0.32 |
| 6 | 1.22 | $9.2\times10^{-3}$ | $8.8\times10^{-3}$ | $3.77\times10^{-3}$ | $3.32\times10^{-3}$ | 0.29 | 0.52 |
| 8 | 1.43 | $(1.1\times10^{-2})$ | $(1.2\times10^{-2})$ | $(2.91\times10^{-3})$ | $(2.31\times10^{-3})$ | (0.13) | (0.30) |
| | $\int F(n)dn$ | 1.00 | 0.940 | 0.972 | 1.01 | - | - |

## Table IV.

### Relation of F(n) and (n).

Date. 22 July, 1952.

Points examined. h' and w' at 3, 4, 5, 6 $c.sec^{-1}$ (8 pairs).

Correlation of log n and log F(n). - 0.965

Regression equation. $\log_{10}F(n) = -2.06 \log_{10}n - 0.45$

Spectrum function. $F(n) = 0.35\ n^{-2.06}$

Date. 11 November, 1952.

Points examined. h' at 3, 4, 5, 6 $c.sec^{-1}$.

w' at 4, 5, 6 $c.sec^{-1}$. (7 pairs)

Correlation of log n and log F(n). - 0.955

Regression equation. $\log_{10}F(n) = -2.03 \log_{10}n - 0.94$

Spectrum function. $F(n) = 0.115\ n^{-2.03}$

Date. 27 January, 1953.

Points examined. u' and v' at 3/2, 2, 5/2, 3, 4, 5, 6 $c.sec^{-1}$. (16 pairs)

Correlation of log n and log F(n). - 0.921

Regression equation. $\log_{10}F(n) = -2.11 \log_{10}n + 0.884$

Spectrum function. $F(n) = 7.65\ n^{-2.11}$

## Table V.

### Terms involved in the turbulent energy equation.

| Date | $\overline{\rho}(h'^2+w'^2)/2$ | $\overline{\rho}\ \overline{h'w'}\ \frac{\partial\overline{u}}{\partial z}$ | $\frac{g\overline{\rho}}{\overline{T}}\ \overline{w'T'}$ | $\frac{\overline{\rho}}{2z}(\overline{w'h'^2}+\overline{w'^3})$ |
|---|---|---|---|---|
| | $erg\ cm^{-3}$ | $erg\ cm^{-3}\ sec^{-1}$ | $erg\ cm^{-3}\ sec^{-1}$ | $erg\ cm^{-3}\ sec^{-1}$ |
| 22 July, '52 | 6 | $+ 2.4 \times 10^{-2}$ | $+ 4 \times 10^{-2}$ | $- 3 \times 10^{-2}$ |
| 11 Nov., '52 | 1 | $+ 0.78 \times 10^{-2}$ | $- 0.7 \times 10^{-2}$ | $- 1 \times 10^{-2}$ |

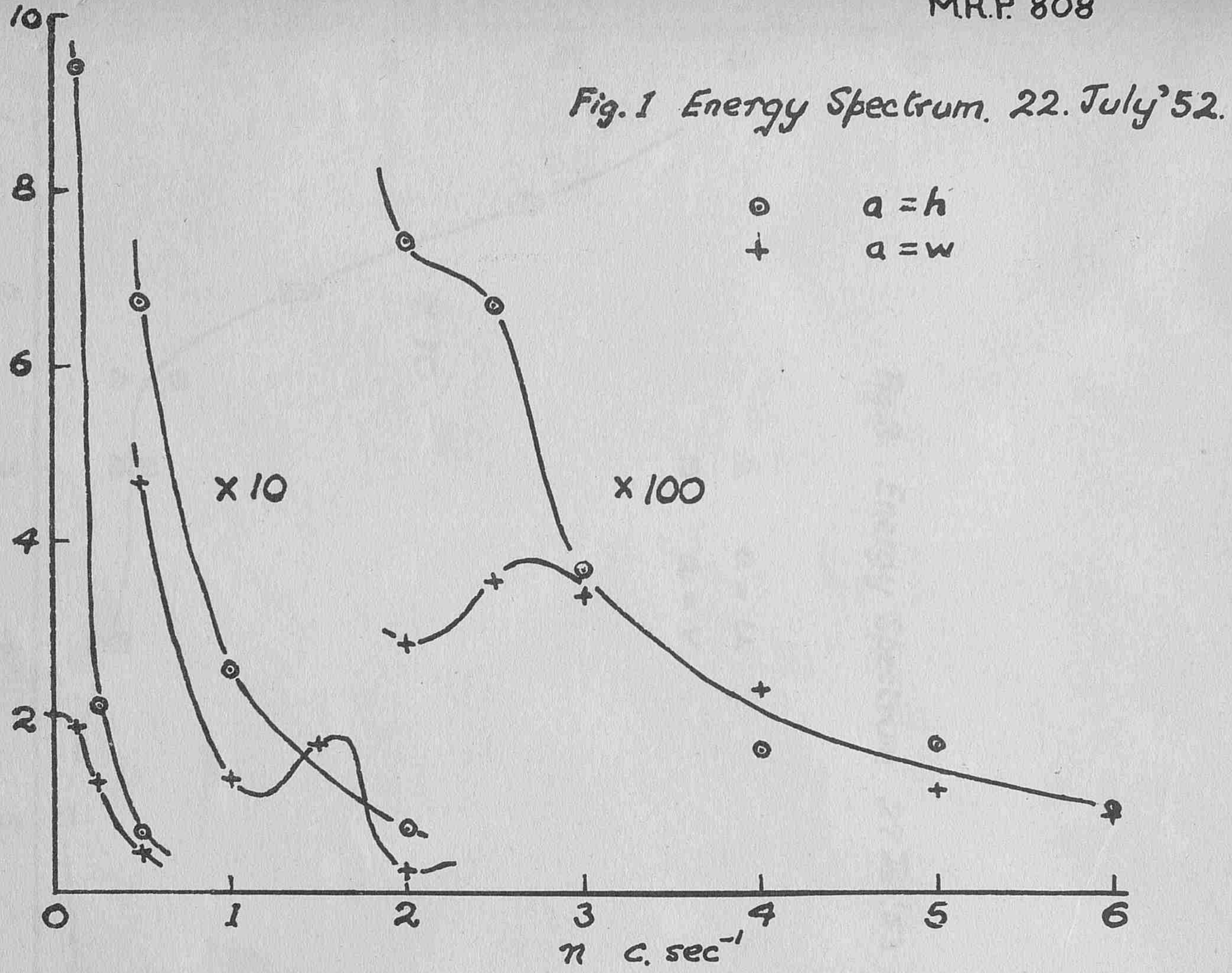

Fig. 1 Energy Spectrum. 22. July '52.

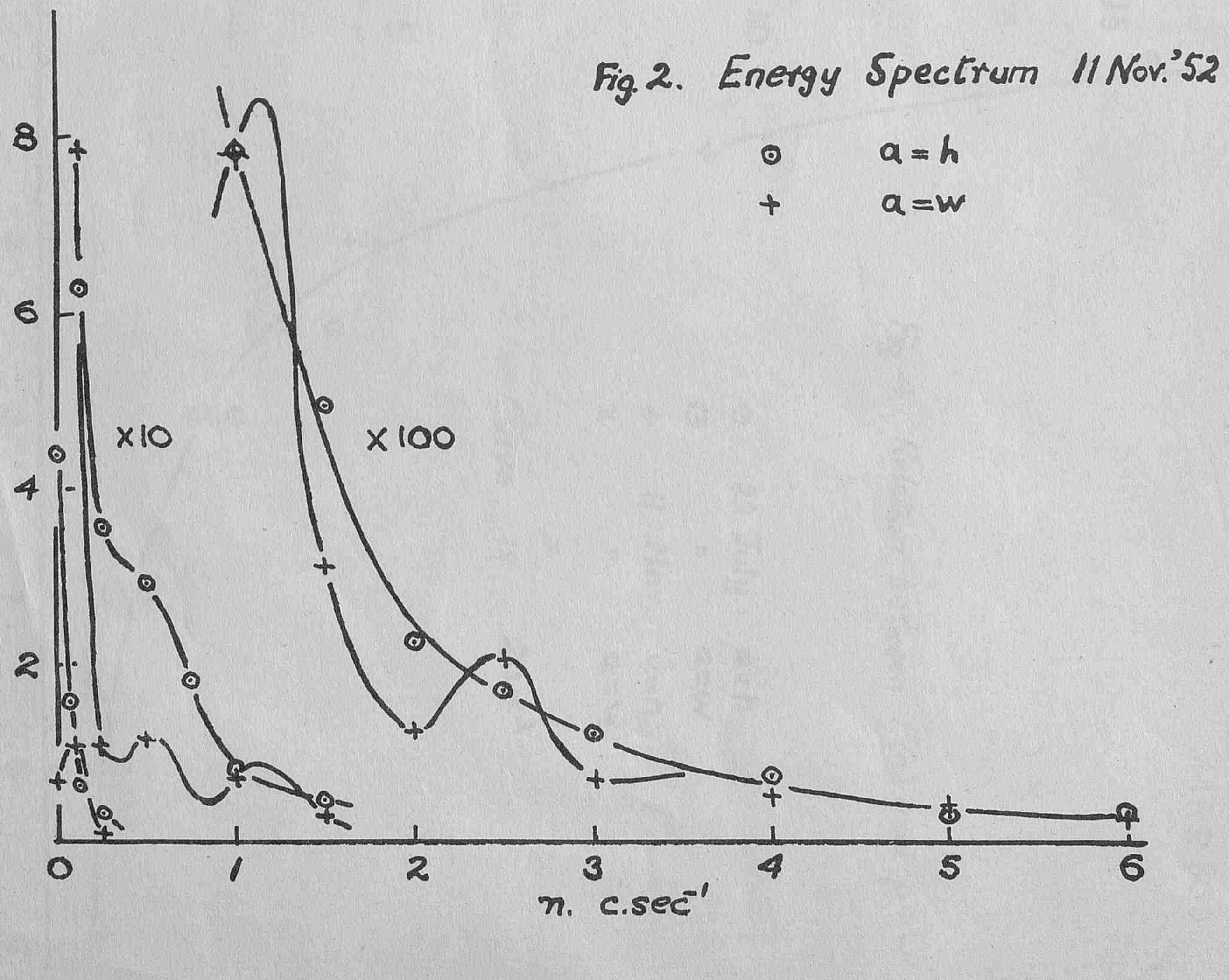

Fig. 2. Energy Spectrum 11 Nov. '52

$\rho\,\overline{a'^2}\,F(n)$ erg cm$^{-3}$/c.sec$^{-1}$

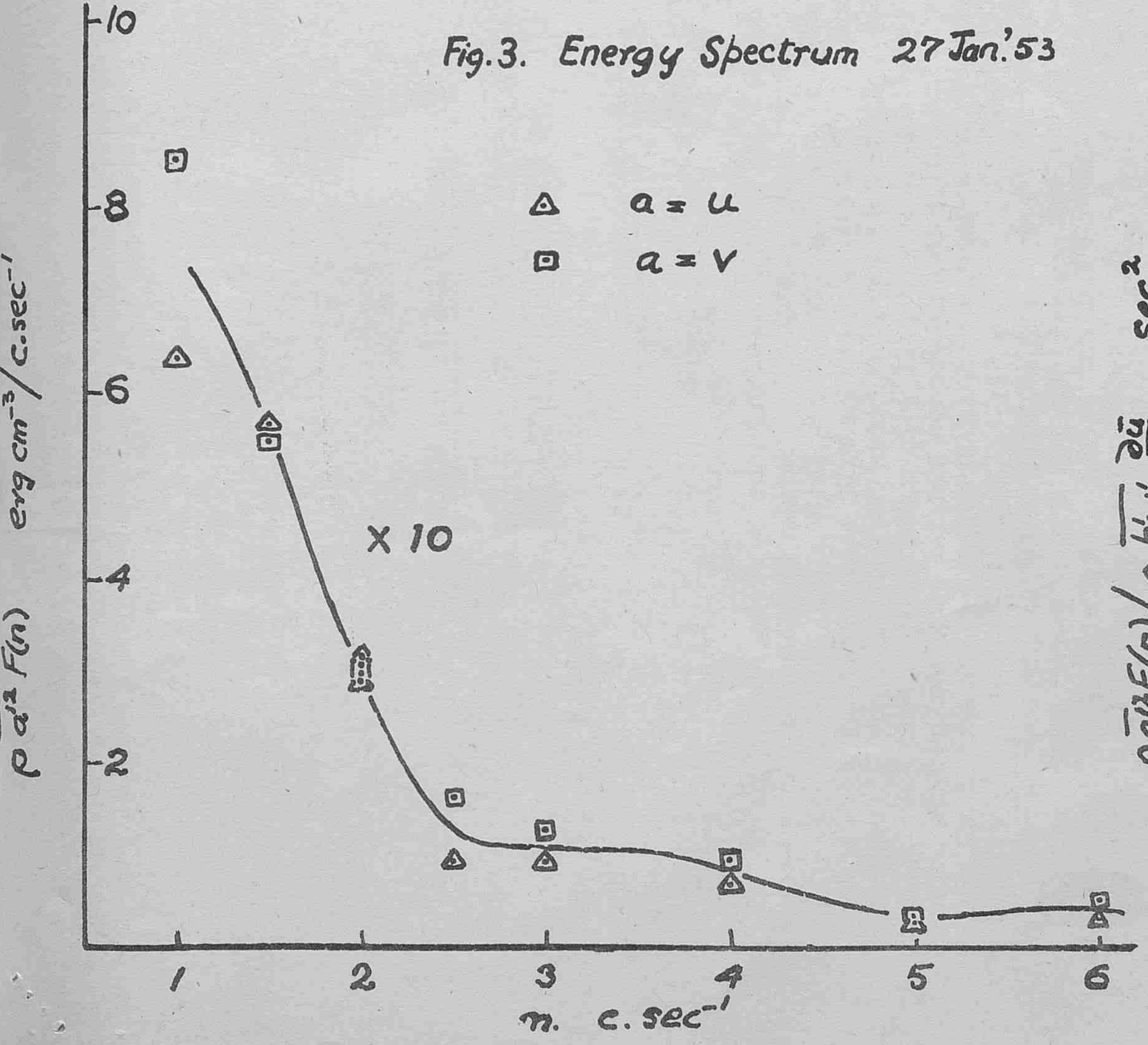

Fig. 3. Energy Spectrum 27 Jan.'53

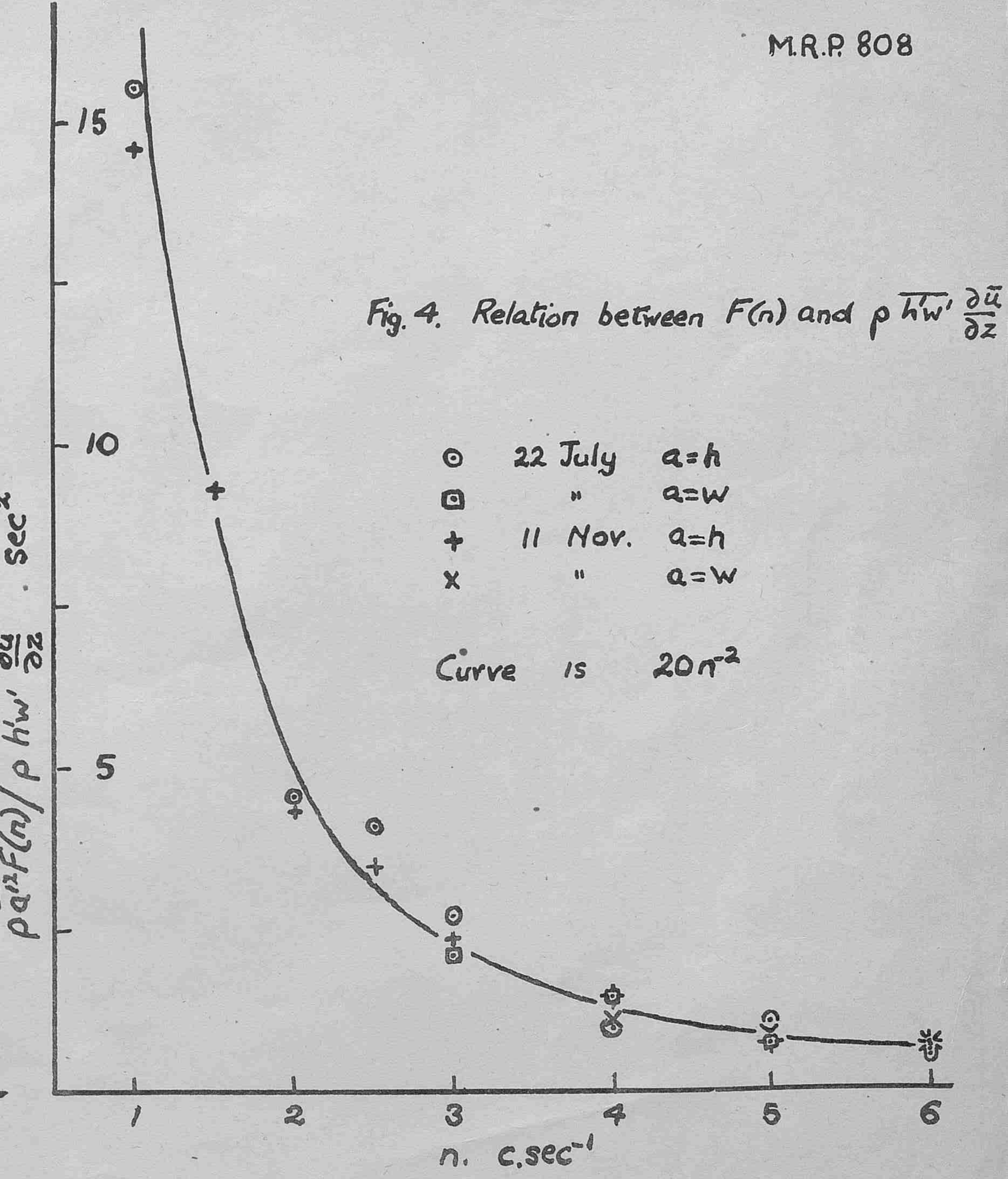

Fig. 4. Relation between F(n) and $\rho\,\overline{h'w'}\,\frac{\partial\bar{u}}{\partial z}$

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AD-44245

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# The eddy diffusion of momentum, water vapour and heat near the ground

by

N.E. Rider (Meteorological Office)

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The experimental data enabled values of the eddy diffusivities for momentum (Km), water vapour (Kv), and heat (Kh) to be obtained on a direct observational basis from the expressions by which the diffusivities are normally defined, and the following conclusions are drawn;

(a) The wind speed and temperature profiles always have the same form and this form is shared on the majority of occasions by the humidity profile

(b) In conditions approximating to adiabatic as defined by small numerical magnitudes of the Richardson Number (Ri) the established laboratory law relating the surface drag to the fluid velocity uz at a height z above a rough surface in turbulent flow

$$u_z = \frac{1}{k}\left(\frac{\tau_o}{\rho}\right)^{1/2} \log\left(\frac{z}{z_o}\right)$$

where $\rho$ is the density of air and zo is the roughness parameter, k being a constant, holds over the present site with k = 0.41 and zo $\simeq$ 0.30 cm. Therefore in these conditions Km may be obtained from observation of the wind profile alone by substitution in the equation

$$Km = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)$$

(c) In conditions other than neutral the above equations no longer hold and it is shown that

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(c) In conditions other than neutral the above equations no longer hold and it is shown that

(i) the equation

$$u_z = \frac{1}{k(1-\beta)}\left(\frac{\tau_o}{\rho}\right)^{1/2}\left[\left(\frac{z}{zo}\right)^{1-\beta} - 1\right]$$

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/which

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which is based on Deacon's generalised wind profile law, $\beta$ being the familiar index, is explicitly true in unstable but not in stable conditions. Km may therefore be obtained from wind profile observations in unstable conditions using the equation

$$Km = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)\left(\frac{z}{z_0}\right)^{2\beta-2}$$

The use of this equation in stable conditions results in a large underestimation of Km.

(ii) the Rossby and Montgomery and Holzman equations

$$Km = k^2 z^2 \left(\frac{\partial u}{\partial z}\right) \Big/ (1 + \sigma Ri)$$

and

$$Km = k^2 z^2 \left(\frac{\partial u}{\partial z}\right) (1 - \sigma Ri)$$

hold in stable conditions with a value of $\sigma$ approximately equal to 8 in the former and 6 in the latter.

(d) The parameters

$$\frac{\tau_0}{\rho z^2 \left(\frac{\partial u}{\partial z}\right)^2} \quad \left[ = \frac{Km}{z^2 \left(\frac{\partial u}{\partial z}\right)} \right]$$

$$\frac{-E_0}{z^2 \left(\frac{\partial \psi}{\partial z}\right)\left(\frac{\partial u}{\partial z}\right)} \quad \left[ = \frac{Kv}{z^2 \left(\frac{\partial u}{\partial z}\right)} \right]$$

are identical irrespective of the stability, so indicating equality of the eddy diffusivities for momentum and vapour in all conditions.

(e) The magnitude of the eddy diffusivity for heat is often approximately equal to that for momentum or vapour but occasions occur when it is larger than the other two. These occasions do not appear to be related to the stability.

/1. Introduction

## 1. Introduction.

Eddy diffusion is of fundamental importance in meteorology, particularly to the meteorology of the microclimatic region. Since the first recognition of the existence of the transfer of heat, momentum and matter in the atmosphere by eddies interest has largely centred on the magnitude of the diffusivities, their dependence on various physical processes and their relationship to one another. A considerable amount of information has accumulated from wind tunnel studies in the laboratory of such factors as aerodynamic drag and evaporation, but, primarily as a result of the difficulties introduced by the nature of natural surfaces and the inability to control any of the many physical processes acting at any time, investigations in the open have tended to lag behind. It has been the usual practice in treating diffusion problems in the atmosphere to assume the equality of at least two of the diffusivities, an assumption that has often enabled the construction of formulae which have proved to be in good agreement with observation. A notable contribution to our knowledge in recent years was made by Pasquill (1949a) who was mainly concerned with the construction and verification of an indirect method of measuring natural evaporation. In the course of this investigation Pasquill was able to demonstrate the equality of the eddy diffusivities for momentum and vapour in neutral conditions by assuming the validity (which he later demonstrated (1950)) of the laboratory law relating the drag exerted on a surface to the fluid velocity above the surface in the lowest layers of the atmosphere. He showed further that in unstable conditions, but not in stable conditions, his observed values of the eddy diffusivity for vapour were in good agreement with values of the eddy diffusivity for momentum when the latter were calculated on the basis of Deacon's (1948) generalised wind profile law. He also questioned the normally accepted identity of the diffusivities for heat and vapour in all conditions of stability. Rider and Robinson (1951) undertook a similar study, the only differences of importance in the observations being the lack of a direct determination of evaporation and a rather more detailed measurement of some radiation components. There were some differences of treatment of the observations and in the results which are of some importance. In particular, Rider and Robinson concluded that it was unlikely that the diffusivities for heat and vapour were not equal in all conditions of stability.

The present work was undertaken to take advantage of a technique for the measurement of the drag exerted on a natural surface by the wind from which values of the eddy diffusivity for momentum could be obtained on a direct observational basis, so permitting a comparison to be made with the diffusivities for water vapour and heat determined simultaneously over the same site. The relation between the form of the wind profile and the aerodynamic drag on the surface can also be studied.

## 2. Site and observations.

### (a) Site and period

The observations were made on the old airfield at R.A.F. Station, Cardington which at the time was in use mainly as a sports field. Throughout the period of observation a circular, more or less level, area of approximately 250 m. diameter was kept mown to a grass length of 2 to 3 cm. and from the centre of this area to a distance of at least 300 m. in an arc from NE through S to WSW there were no obstacles to the flow of air. In an arc from WSW to NNE at a distance of 250 to 300 m. from the centre of the close mown area two large buildings (the airship hangers) were situated. To avoid any possibility of unrepresentative results arising from a modification of the character of the air flow by these buildings observations were only made when wind directions within the former arc were experienced. Outside the close mown area there was sometimes a change in grass length, the length approaching 10 cm. at times, and it was feared that the change in surface roughness characteristics sometimes present at about 150 m. from the apparatus would have some undesired effect on the form of the profiles at heights approaching 2 m. In the event no such effects were detected. It was not possible to have a larger area close mown at frequent intervals.

Observations were made in the period July to November 1952. In all fifty one separate 30 min. observations were taken but it was not found possible to observe all the appropriate quantities on every occasion. Owing to a lack of observers, in the first series of observations (Nos. 1 - 26, see table I) only the wind speed and temperature profiles were measured together with a determination of the drag exerted on the surface. In the second series (Nos. 27 - 51, see table II) it was possible to add to these observations that of the humidity profile, a direct determination of the natural evaporation, and in addition, when conditions were suitable, measurements of the gain or loss of heat in the ground and the long and short wave radiation components. During one observation in the first series and a few in the second series no measurement of surface drag was made; this occurred when the wind direction was unusually variable and changed by as much as 180° during the course of the runs.

## (b) Air temperature and humidity

These were measured with an installation of dry and wet bulb aspirated thermoelectric thermometers mounted on a mast specially constructed for the purpose. The installation has been described in detail by Pasquill (1949b) and the mast and thermometers were subjected to all the tests he recommends. The psychrometer units were mounted at six heights, namely 25, 37.5, 50, 100, 150, and 200 cm. above the surface and a ventilation rate of about 3 m. sec.$^{-1}$ was provided by the aspiration fans. The method of wiring and switching was such that the temperature of the 2 m. dry bulb was determined to 0.1°F with respect to a flask of melting ice. The dry bulb temperatures at the other five heights were then found as differences to 0.01°F from the 2 m. dry bulb. On occasions when the humidity profile was observed the wet bulb depression at each height was then obtained to 0.01°F. In a 30 min. observation period readings were taken for two 12 min. periods separated by an interval of 4 min. These 4 min. were occupied in interchanging the psychrometer units up and down the mast to reduce as far as possible any instrumental error on the final profiles. This procedure lengthened the analytical work but was thought necessary as even the most careful adjustment and calibration does not produce six perfectly matched psychrometer units. A period of about 1 min. was required to take the 12 galvanometer readings necessary for one determination of the temperature and wet bulb depression at each height so that 24 or so determinations of each were available in each observation. When only one observer was available to read the galvanometer and record the deflections the wet bulb depressions were not taken (observations 1 to 26) and rather more than 24 separate observations of the temperature at each height were obtained. In constructing the mean temperature profile the individual observations were meaned. For the humidity profile the wet bulb depression associated with each determination of the temperature at each height was used to calculate the vapour pressure on the basis of the usual psychrometric formula (M.O.265) and the values of the saturated vapour pressure (interpolated as necessary) taken from the Smithsonian Tables. The individual absolute humidities at each height were then calculated and the 24 or so values meaned.

## (c) Wind

Wind speed measurements were made with sensitive cup anemometers similar to those described by Sheppard (1940). On all occasions eight were exposed at heights of 15, 25, 37.5, 50, 75, 100, 150, 200 cm. It was not possible to mount all on one mast so it was expected that a certain amount of scatter would be present in the results owing to the small scale undulations of the natural surface, the height of each anemometer being set with reference to the ground at the foot of its mounting. (For a more detailed account of this difficulty which is always present when a number of anemometers are exposed over a natural surface see Rider 1951, para. 10.) The anemometers were allowed to run for two periods of 10 min. (12 min. in those few observations in which the drag plate apparatus was not used) separated by a 6 min. interval in each observation and during this interval of 6 min. between the two runs the instruments were interchanged, here again to reduce the influence of any instrumental error on the mean profile which was constructed from the mean of the two wind speed determinations at each height. During the course of the investigation the anemometers were calibrated in a wind tunnel.

/(d)

### (d) Aerodynamic drag

The drag plates available were generally similar in design to those used previously by the writer (1951). Six were constructed at Kew Observatory and the writer is indebted to the Superintendent and Staff for this service. Each apparatus consisted essentially of two vertical sided spun aluminium pans having diameters of 18 in. and 19 in. respectively. The smaller, slightly more shallow, pan was provided with a false bottom on which a sample of the natural turf surface some 2 to 3 in. thick could be supported. A circular cutting ring was made to aid the extraction of surface samples. The inner pan was provided with two rigid links at opposite ends of a diameter. One supported a rigid brass clamp, the other enabled a 30 in. long light duralumin arm with an air damping plate at its centre and a syphon type pen at its far end, to be attached rigidly to the inner pan. The inner, smaller, pan carrying the sample of the surface floated in water contained in the larger pan and the pen at the far end of the arm rested on a chart over a drum mounted on a horizontal axis at right angles to the axis of symmetry of the apparatus. This drum was driven by a synchronous motor to rotate once in 30 min. and the drum diameter was sufficient to provide a chart travel of about 1.5 cm. per minute. Control of the movement of the inner pan in the horizontal was provided by means of a vertically mounted phosphor bronze suspension held by suitable attachments to the outer side of the larger pan. The brass clamp attached to the smaller pan enabled the latter to be attached to the centre of this suspension. Arrangements were provided for tensioning the suspension and adjusting the zero, and various suspensions could be inserted as required. In the field it was found that two suspensions of different cross section were sufficient to allow reasonable deflections over the range of wind speeds experienced. The sensitivity of the drag plates varied slightly from one to another and with the tension applied to the suspensions but for the thinner suspension (diam. .048 in.) the overall sensitivity was about 3 cm. per dyne cm.$^{-2}$ and for the thicker suspension (diam. .065 in.) the corresponding figure was approximately 1 cm. Calibration of the plates used on any day was performed before and after the day's observations, in the laboratory.

In the field sufficient turf and soil were removed to allow the outer pan to be inserted to below the level of its top rim. A box containing the pen arm was let into the surface together with another box containing the drum and its driving mechanism. These boxes were covered with perspex lids. In situ the only parts of the apparatus which projected above the air / earth interface were the top of the torsion head and the links to the pen arm and controlling suspension from the inner pan. All these projections were within the grass length and the links were screened from the horizontal wind by faired plates. It is likely that some small error was present in the observed surface drags as a result of wind pressure on the small side areas of the links. No correction was possible but any error here was certainly small when compared with the uncertainty introduced by the non-representative nature of the surface sample and its almost certainly incorrect level on any one occasion, which will be referred to below.

Six sets of apparatus were available and were sited in the ground with their axes of symmetry as tangents to a circle of about 8 m. radius spaced 40° apart over the arc from which unobstructed winds were available. A light bi-directional wind vane was mounted on a mast 1.25 m. high at the centre of this circle and during an observation the two plates over and between which the wind was expected to blow to the vane were allowed to record. Covers were provided in order to obtain the zero positions of the plates at the start and finish of each run, and the procedure was to take a 30 min. trace from the two plates of which the first and last minutes were occupied by zero lines, the covers being removed and replaced at 1 and 29 min. respectively. The parts of the traces on which computations were based occupied the periods 2 to 12 min. and 18 to 28 min., the two periods of anemometer operation. Temperature and humidity measurements were made in the periods 1 to 13 and 17 to 29 min. Four 3 min. traces were provided by the vane, two in each period of drag plate operation. From the two drag plate records the mean deflection and hence the force per unit area of exposed sample surface during each 5 min. period which had to be associated with each of the four determinations of wind direction was obtained with the aid of a planimeter on each 5 min. 'area'. Two determinations of the force per unit area were thus available for each 5 min. period. The wind direction records enabled the angle between the 5 min. mean winds and the perpendicular to the axes of symmetry of the drag plates to be calculated, and from the force and this angle the 5 min. mean drags were obtained. In calculating the mean value of the drag for the whole period of an observation which appears in tables I and II, a selection from the two determinations in each of the four 5 min. periods was made, the value used being provided by the plate for which the angle mentioned above was the smaller. This angle was never larger than 20° for the 5 min. determinations chosen.

It soon became apparent that the values of the surface drag provided by both plates in each period were not the same and differed by as much as 30%. This is not surprising when it is remembered that, as well as the difficulty of choosing and cutting two 18 in. diameter samples of the surface which were identical aerodynamically and were both typical of the surface of the experimental site as a whole, the level of the inner pans carrying the samples had to be adjusted to expose the samples at their natural levels. This was found to be a great difficulty in practice and in considering the results one must bear in mind that any one value of the surface drag given in tables I and II may be in error by an amount up to at least 30%. Between each observation the pans were relevelled and every few days new samples of the surface were cut and inserted in the plates. In this way it was hoped to produce a result which would at least have statistical significance and it will be seen later that this has been the case.

## (c) The direct measurement of evaporation

This has been attempted in the second series of observations by the use of simple soil evaporimeters, the technique being to extract undisturbed cores of the turf and soil using a suitable cutter, place them in tufnol containers, and replace them in the containers in their natural positions in the ground. To facilitate easy removal for weighing, the tufnol containers holding the cores are placed in other, slightly larger, containers let into the surface. The technique used follows that used by Pasquill (1949a) but the soil type on the present site was not the same as that on the site of Pasquill's experiments and some explanation and justification of the method is necessary.

Rates of evaporation obtained from simple evaporimeters of this type are liable to errors which arise as a result of the isolation of the soil cores from their surroundings. The isolation from lower layers may cut off the supply of water which may normally be transported upward to the core level, so that after some time the amount of water available in the core may be less than that available in the surface generally. Further, the breaking of roots will reduce water loss by transpiration. If the soil is heterogeneous in particle size the water content will vary from place to place and there will be a difference between the rate of evaporation from the small area of an evaporimeter and that for the site as a whole. The temperature distribution in the isolated core cannot be expected to remain the same as that in the undisturbed ground and this may have some influence on the measured evaporation. The effect of isolation in depth may be investigated by using evaporimeters of different depth at the same time, and any errors due to the unrepresentative nature as regards water content of a single core may be minimised by using the mean result from a number scattered over the site. No correction for, or estimation of, the influence of non-representative temperatures in the cores seems possible; we can only use containers of low thermal conductivity and hope that the effect will be small.

In order to examine the soil structure on the site a hole was dug to a depth of 3 ft. to the NW of the apparatus. From the surface down to 16 in. to 18 in. the soil was a gravel loam and below this was a layer of gault clay.

The turf roots largely ceased at a depth of 3 in. but an occasional root, usually that of a plantain, extended to 6 in. or more. An examination of the walls of the hole showed that there were no layers which appeared to be wetter than the first few inches, and throughout the period of observation no water collected in the bottom of the hole. The variation of water content with depth down to 8 in. was measured on a number of occasions for the calculation of the specific heat of the soil material (see below) and on every occasion the water content decreased with depth. We may therefore assume that there was no general upward movement of water from depths below that of the deepest evaporimeter used, and that over the period of the observations the surface layers were effectively disconnected from lower layers in so far as water movement was concerned. In placing the drag plates in position a considerable amount of surface soil was removed to a depth of 8 in. to 10 in. and it was possible to examine the variation in soil water content qualitatively over a fairly large area. The variation seen was very marked and this accounts for the large differences in evaporation often obtained from the various evaporimeters exposed together. The apparent density of the core material varied considerably from evaporimeter to evaporimeter and it was noticeable that the heaviest cores usually showed the greatest water losses. To provide support for the contention that the isolation of the cores in depth produced no serious error in the observed water losses, a preliminary experiment was conducted in which six evaporimeters, type A, having a depth and diameter of 4 in. and six, type B, having a depth of 8 in. and diameter of $2\frac{3}{8}$ in. were exposed at random over the site and the water losses from each measured over three separate intervals of an hour. The results are given in table IV below. It was not possible to cut satisfactory cores of 4 in. diameter and depth 8 in. in the gravel loam and in any case such cores were not within the capacity of the balance, which was semi-automatic and could be read to 0.01 g. and was consistent to 0.02 g. The twelve evaporimeters were set out in a time of about 90 min. and fresh cores were put into each evaporimeter for each hour's readings.

Table IV

Water losses per $cm.^2$ of surface from six type A evaporimeters and six type B evaporimeters expressed as percentages of the loss from one type A evaporimeter on 16th September 1952 (a) 1100-1200, (b) 1400-1500, (c) 1700-1800 G.M.T.

| | Type A | | | | | | | Type B | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | Mean | 1 | 2 | 3 | 4 | 5 | 6 | Mean |
| (a) | 100 | 62 | 112 | 105 | 77 | 93 | 92 | 87 | 132 | 96 | 90 | 77 | 50 | 89 |
| (b) | 100 | 137 | 93 | 67 | 99 | 113 | 102 | 109 | 97 | 127 | 96 | 103 | 129 | 110 |
| (c) | 100 | 147 | 107 | 184 | 141 | 131 | 135 | 173 | 81 | 156 | 98 | 131 | 81 | 120 |

The results show the scatter that was always present in the performance of the evaporimeters, but no systematic effect of evaporimeter depth is evident. Since the type A evaporimeters were more convenient, these were used in the general programme of observations. The rates of evaporation shown in table II are the mean result from six type A evaporimeters for which new cores were cut and inserted on every day of observation to eliminate, as far as possible, any influence of core age and consistent errors due to non-representative selection of samples on any one day.

During a few observations the mean result from the six evaporimeters showed an increase of weight over the period of the observation indicating condensation at the surface. To what extent the evaporimeters used provide a satisfactory measure of dew fall cannot be ascertained. However, we may note that the results in such cases are in keeping with the general run and there seems to be no reason to reject them out of hand. In a few observations the rate of evaporation was very small, and for these occasions no value of the rate of evaporation, Eo, is given in table II.

/(f)

### (f) Flux of heat in the ground

Earth temperatures were obtained by a thermoelectric earth thermometer having 'bulbs' at depths of $0, \frac{1}{2}, 1, 2, 4, 6, 8, 12$, and 16 in. below the surface. The instrument has been described elsewhere (Pasquill 1949a). Two sets of readings were taken during each observation, one at the start and the other at the end, and it is assumed that any temperature changes were linear in time. Little error can arise from this assumption as soil temperatures are known to change slowly. Another requirement was a knowledge of the specific heat of the soil material. This was found by taking at the start of the second series of observations soil cores in three places, with a suitable extractor, down to a depth of 8 in. and determining, by the familiar method of mixtures, the specific heat of the dry soil material at various levels. There was considerable scatter in the determinations but no systematic variation with depth and the value $0.20 \pm .05$ was obtained. The value 0.20 has been used in the calculations. On each day of heat balance observation three more cores were cut and the water content in the layers 0-1, 1-2, 2-4, and 4 - 8 in. was determined by weighing before and after drying. It was then a simple matter to calculate the actual specific heat in the various layers on every occasion, and this, together with the temperature readings, was all that was required to find the gain or loss of heat in the ground over the period of an observation. A detailed account of the procedure is given by Pasquill (1949a).

### (g) The radiative flux of heat

A Moll solarimeter calibrated at Kew Observatory was used to measure the short wave radiative flux. Continuous recording was not possible and readings of the solarimeter output were taken every 2 min. From time to time the instrument was inverted to obtain a measure of the percentage of the incident radiation reflected at the surface.

The long wave, atmospheric, radiation from the ground and sky was estimated by the method described by Robinson (1947), the only observation required on the site being the reading of a spirit in glass thermometer resting on the surface. In constructing the water content/temperature curve on the radiation chart the readings of temperature and humidity up to 2 m., taken on the site, were used. For the curve above this, the results of radio-sonde ascents at Larkhill and Hemsby were employed.

## 3. The profiles and vertical gradients of wind, temperature and humidity.

The observed values of the elements from which the profiles have been constructed are set out in tables I and II. In the computations which are to follow we shall require the vertical gradients of the elements at various heights and it is therefore necessary to consider the forms of the profiles and the method to be used in deducing the gradients. Pasquill (1949a) in deriving his gradients assigned, for the majority of his observations, the same functional form to the three profiles and obtained his gradients arithmetically whereas Rider and Robinson (1951) preferred to make no assumptions as to the form of the profiles and obtained gradients by simple tangent drawing to the best smooth freehand curve that could be constructed through the individual measured values when the latter were plotted against a linear height scale. Rider and Robinson's method has the advantage that no functional form has to be chosen and any results which arise from the determined gradients cannot be attributed to this choice. These investigators, on determining their gradients, found that in many cases the ratios of the gradients of any two elements were approximately constant with height and this led them to plot the measured values of the elements on one curve after the scales of the elements had been adjusted so that the best freehand profiles drawn for each coincided at two heights. They were able to show that in the majority of their observations the profiles had the same form i.e. one curve could be drawn on the combined plot to represent the profile of each of the three elements. However, exceptions occurred when this could not be done. In the present work the measured values of the elements were plotted against a linear height scale and the best

/freehand

freehand curve was constructed through the points. In almost all cases this could be done with confidence although an individual measurement here and there appeared to be in error. The scales were then adjusted so that the profiles drawn by hand would coincide at 37.5 and 150 cm. and the wind, temperature, and humidity measurements were then plotted on one curve using the adjusted scales. In the first series of observations, Nos. 1 - 26, temperature and wind measurements only were available and it was found that in every instance one curve could be drawn on the combined plot to represent the most likely profile of both temperature and wind. Fig. 1 shows the superposed plots for observations 14 and 19, the first a fairly typical example, the second the worst in the series. In the second series of observations, Nos. 27 - 51, which included measurements of humidity, it was found that the wind, temperature, and humidity profiles could be represented by one curve after suitable adjustment of the element scales in nineteen of the twenty-five observations. In five cases (Nos. 27, 30, 37, 47, and 48) the humidity profile appeared to have a different functional form and could not be represented by the same curve as the wind and temperature. However, it must be noted that in only one of these five observations (No. 37) did the individual humidity points show a departure from the superposed wind and temperature profile which was outside the accuracy that could be expected in the humidity determinations. Nevertheless, the departures of the humidity points were systematic in all the five observations. Fig.2a shows a typical example in observation 30 and fig.2b shows a plot for observation 37. Two examples, observations 31 and 49, where one curve was drawn on the superposed plot to represent the most likely profiles of the three elements are illustrated in fig.3. In observation 50 the humidity profile was vertical within the limits of measurement and no conclusion regarding its form can be reached. The observations were made in all conditions of wind speed, temperature, stability, sky, etc. and it appears permissible to conclude that the wind and temperature profiles have the same functional form and that this form is shared by the humidity profile on the majority of occasions. Exceptions appear to occur in the case of the humidity profile which, as far as can be ascertained at present, are not associated with any particular conditions.

At this stage the method used to obtain the gradients of wind, temperature and humidity must be explained. For the first group of observations the wind and temperature profiles ( $\partial u/\partial z$ and $\partial T/\partial z$ ) were measured from tangents drawn to the superposed profiles and the same has been done for the wind, temperature and humidity ( $\partial \psi/\partial z$ ) gradients in the nineteen observations in the second series for which it was found that one curve, after suitable adjustment of the element scales, could be drawn to represent the three profiles. For the other observations in the second series in which the humidity profile appeared to have a functional form different from that of the temperature and wind, the humidity gradients were found by tangent drawing to the best curve that could be constructed through the humidity points, the gradients of the other two elements being found from their common profile. In observation 50 the humidity gradients were indeterminate. Values of the gradients so found are set out in tables I and II. The temperature and humidity gradients are given to $1 \times 10^{-4}$ °C cm.$^{-1}$ and $1 \times 10^{-10}$ g cm.$^{-4}$ respectively and in general when the gradients were found to have a magnitude less than ten times these values they have not been given as an error of more than 10% has a large influence on the magnitudes of the quantities which have been deduced from the gradients, particularly when the gradients are multiplied or raised to a power. There is also a considerable uncertainty in drawing the tangents in such cases.

The reliability of gradients determined in this way has often been questioned and it is very desirable to arrive at some estimate of the degree to which the freehand profiles and gradients can be reproduced. After the whole series of observations had been treated in the manner described above, one observation, No. 14, was chosen and three observers plotted the measured wind speeds and temperatures, drew the freehand profiles independently, and constructed the superposed profile on the basis of the individually determined 'smooth' values of the elements at 37.5 and 150 cm. Each observer then obtained gradients from his own superposed profile at the heights of 37.5, 75, and 150 cm. The results for the temperature gradients are given below. (The wind speed gradients differ from the temperature gradients by a constant factor.)

/Observation No. 14.

Observation No. 14.

| Observer | $\frac{(\partial T)}{(\partial z)} \times 10^{-4}$ °C cm.$^{-1}$ at z cm. | | |
|---|---|---|---|
| | z = 150 | z = 75 | z = 37.5 |
| N.E.R. | 21 | 82 | 196 |
| E.L.E. | 22 | 76 | 187 |
| P.H.L. | 17 | 71 | 197 |
| Means | 20 ± 2 | 76 ± 4 | 194 ± 6 |
| Original determination by N.E.R. as in table I. | 22 | 75 | 209 |

This suggests that the gradients given in the main tables are reliable to at least 10%. It must be noted that the major part of the uncertainty is introduced in the construction of the profiles rather than in the construction of tangents to a given profile. The three observers were able to agree to the values given for the gradients to within $2 \times 10^{-4}$ °C cm.$^{-1}$ when checking one another's gradients in the case given above. In the arithmetical method used by Pasquill errors in the measurement of temperatures differences of the order of 0.02°C would give rise to errors in the deduced gradients comparable to the uncertainty inherent in the method used in the present work.

As the functional forms of the profiles of the elements appeared to be the same for the majority of occasions we must expect a law, similar to the Deacon (1948) generalised wind profile law, to hold for the temperature profile and the humidity profile on most occasions. Deacon, after a long series of observations, proposed the wind profile law

$$\frac{\partial u}{\partial z} = a\, z^{-\beta} \qquad 1.$$

where $\beta$ takes values $< 1$, $= 1$, and $> 1$ in stable, neutral, and unstable conditions respectively. In neutral conditions the law reduces to the familiar u, log z form. Studies by Deacon and others have resulted in the recognition that the form of the wind profile is controlled by the magnitude of the wind speed as well as by the temperature gradient. This is to be expected as the wind speed is a measure of the shearing force which opposes any modification of the air flow by buoyancy forces of which the temperature gradient is a measure. The Richardson Number, Ri, which is defined by

$$Ri = \frac{g}{T}\left\{\frac{\partial T}{\partial z} + \Gamma\right\} \Big/ \left(\frac{\partial u}{\partial z}\right)^2$$

where g is the acceleration due to gravity, T is the absolute temperature and $\Gamma$ is the dry adiabatic lapse rate ($- 1 \times 10^{-4}$ °C cm.$^{-1}$) and takes positive and negative values in stable and unstable conditions respectively, has proved to be a convenient indicator of the state of the atmosphere for this purpose. Its magnitude has been calculated at the three heights 37.5, 75 and 150 cm. for the present observations where possible and the values are set out in tables I and II.

A method which has been used to study the variation of the profiles with stability in the past is to plot values of the parameter

$$\frac{S_{z_1} - S_z}{S_{z_1} - S_{z_2}}$$

where Sz represents the wind speed, temperature, or humidity as desired at the height z, z1 and z2 being two fixed heights, against the height on a logarithmic scale. The present measurements were investigated in this way and it was found that, in general, the S, log z plots were concave to the log z axis in unstable conditions, convex to it in stable conditions, and showed an approximately linear variation of S with log z in near neutral conditions for the three elements. Exceptions occurred which are thought to be due to a large extent to experimental errors in magnitudes of the elements, particularly at z1 and z2 as the magnitude of (Sz1 - Sz2) has a large influence on the shape of the plot. An illustration of the general features of these plots is given in fig. 4 in which the mean values of the parameter for wind speed, temperature, and humidity observations in three unstable (Nos. 31, 32, and 33), three near neutral (Nos. 28, 38, and 46), and three stable (Nos. 29, 40, and 42) cases with z1 = 25 cm. and z2 = 50 cm. have been plotted against the height on a logarithmic scale. The wind speed measurements at 15 cm. have not been used here.

## 4. The eddy diffusivity for momentum.

In specifying the eddy diffusivity for momentum at the height z, Kmz, it has been the custom (see e.g. Calder 1949) to employ the well established laboratory empirical law relating the drag exerted on an artificially roughened surface to the fluid velocity at a height z above the surface, uz. This law may be written in the form

$$u_z = \frac{1}{k}\left(\frac{\tau_0}{\rho}\right)^{1/2}\log_e\left(\frac{z}{z_0}\right) \qquad 2.$$

where k is a constant which has the value 0.40, $\tau_0$ is the drag on the surface, $\rho$ is the density of air which will be given the value $1.2 \times 10^{-3}$ g. cm.$^{-3}$ here, and zo is the roughness parameter of the surface and has the dimensions of length and a different value for every type of surface. It is sometimes necessary to replace z by (z - d) in meteorological applications when the height of the roughness elements is comparable with z which is measured from the solid air/earth boundary. d is known as the zero displacement of the surface. It is apparent that equation 2 can only hold in the atmosphere when there is a logarithmic variation of wind speed with height and the necessary functional form of the profile only occurs in near neutral conditions as specified by small numerical values of the Richardson Number. Sheppard (1947), Pasquill (1950) and Rider (1951) have all shown that equation 2 holds in the lower atmosphere in near neutral conditions with values of k approximately equal to the laboratory determined value. The observations of wind profile and surface drag presented here afford a further opportunity to explore the validity of this equation and to obtain values for k and zo which will be required later. Of the 51 observations 14 were conducted in near neutral conditions as specified by values of the Richardson Number at 75 cm., Ri 75, $\leqslant \pm 10 \times 10^{-3}$. The details of these observations have been taken from the main tables and set out in table V which contains values of the ratio uz/u75 and of $u_{75}/\tau_0^{1/2}$ . Plots of the individual observations of wind speed ratio show some departures from a smooth u, log z form and these departures arise in part from the difficulty in setting the heights of the anemometers relative to one another (see para. 2(c)) and to the departure from neutral conditions. However, a plot of the mean values of the ratio against the height on a logarithmic scale shows a good approach to a straight line with some suggestion of curvature concave to the log z axis. This may be the result of the existence of a small zero displacement d due to the grass length but having regard to the mean value of the Richardson Number and to the accuracy of setting of the anemometer heights, the curvature is thought to be insufficient to clearly indicate the need to introduce a value for d in this case. We may note that the u15/u75 point would lie on the straight line drawn through the uz/u75 points shown in fig.5 if the point were plotted at z = 14 cm. The maximum value that could be assigned to d would be about 1 cm. which is small compared with the majority of the heights of wind speed measurement.

The equation to the line of fig.5 is

$$uz/u75 = (\log z + 0.49)/2.38$$

which, when compared with equation 2 written in the form

$$\frac{uz}{u75} = \frac{2.303}{ku75}\left(\frac{\tau_0}{\rho}\right)^{1/2}\left[\log_{10} z - \log_{10} z_0\right]$$

yields a value for zo = .32 cm.

The mean value of $u75/(\tau_0)^{1/2}$ is 385 and if this is associated with the condition

$$\frac{2.303}{ku75}\left(\frac{\tau_0}{\rho}\right)^{1/2} = \frac{1}{2.38}$$

we obtain k = 0.41
which is in good agreement with the laboratory determined value. Equation 2 therefore holds over the present site in neutral conditions with k=0.41 and zo=0.32 cm.

Now Kmz is defined by

$$Kmz = \frac{-Mz}{\rho\left(\frac{\partial u}{\partial z}\right)_z}$$

where Mz is the vertical flux of momentum at the height z. If it is now assumed that momentum is conserved in the turbulent flow we may write

$$Kmz = \frac{\tau_0}{\rho\left(\frac{\partial u}{\partial z}\right)_z}$$

and this, together with equation 2, leads to the equation

$$Kmz = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)_z \qquad 3.$$

and provides us with a valid method of finding values for Kmz in neutral conditions from observation of the wind profile alone.

In stable and unstable conditions this method of finding Kmz cannot be used except possibly at small values of z where the departure from neutral conditions and hence from the u, log z form of the wind profile as indicated by the magnitude of the Richardson Number is small on most occasions. In non-neutral conditions generally, Deacon's generalised wind profile law provides a possible method of specifying Km from wind profile observations. By integrating equation 1 and making use of his observation that the influence of the stability effect on the departure from the u, log z form decreases as the surface is approached, Deacon obtained the equation

$$uz = \frac{1}{k(1-\beta)} \left(\frac{\tau_o}{\rho}\right)^{1/2} \left[\left(\frac{z}{z_o}\right)^{1-\beta} - 1\right] \qquad 4.$$

which is analogous to equation 2 and reduces to it for small values of z and for values of $\beta$ not differing greatly from unity. A difficulty experienced with this equation was that the values of zo determined from it showed a variation with stability which Deacon considered to be unlikely except perhaps at extreme stability and he suggested that the apparent variation was probably due to a small variation of $\beta$ with height. To avoid the difficulty Deacon made the assumption that zo was independent of stability and that its value was satisfactorily defined by equation 2 and wind profile measurements in neutral conditions. Using the value of zo so obtained and wind profile measurements over the same surface in non-neutral conditions equation 4 was then used to calculate a mean value for $\beta$ for the layer concerned.

Assuming the validity of equation 4 Kmz is given by

$$Kmz = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)_z \left(\frac{z}{z_o}\right)^{2\beta-2} \qquad 5.$$

or we may write

$$\frac{Kmz}{z^2 \left(\frac{\partial u}{\partial z}\right)_z} = k^2 \left(\frac{z}{z_o}\right)^{2\beta-2}$$

which gives

$$k^2 \left(\frac{z}{z_o}\right)^{2\beta-2} = \frac{\tau_o}{\rho z^2 \left(\frac{\partial u}{\partial z}\right)_z^2} \qquad 6.$$

from the expression defining Kmz.

From our results values of $\tau_o/\rho z^2 \left(\frac{\partial u}{\partial z}\right)_z^2$ are readily available and have been set out in tables I and II, and it only remains to obtain values of $\beta$ from our wind profile measurements before the validity of equation 6, and hence of equation 4, may be investigated. There are a number of ways in which values of $\beta$ may be obtained but it is an unfortunate fact that values of $\beta$ found on the basis of single observations of the wind profile are very sensitive to errors in the wind speed measurements and it is impossible to derive satisfactory values of $\beta$ from single observations using the anemometers that were available. We are therefore obliged to use a method in which the observations can be smoothed and this in turn leads to a smooth variation in the values of $\beta$ so found. The measured values of u150/u37.5 were plotted against the Richardson Number at 75 cm., Ri75. There was some scatter but it was evident that a smooth curve could be drawn through the points and in order to aid the construction of this curve the mean values of the ratio in ranges of Ri75, together with the mean value of Ri75 in each range, were obtained and plotted. This plot is shown in fig. 6 in which the number attached to each point is the number of separate observations represented by the point. The smooth curve drawn through the points is considered to be fairly well defined between the limits Ri75 x $10^3$ = +100 to −300, and indicates a value of u150/u37.5 = 1.285 at Ri75 = 0 which together with equation 2 gives a value of zo = 0.30 cm. which has to be compared with the value of 0.32 found previously. The degree of agreement is considered to be satisfactory as a change in u150/u37.5 at Ri75 = 0 of approximately ± .005 is sufficient to change zo by ± .02 cm.

From equation 4 we have

$$\frac{u\ 150}{u\ 37.5} = \frac{(150^{1-\beta} - z_o^{1-\beta})}{(37.5^{1-\beta} - z_o^{1-\beta})}$$

and using the value of zo = .30 cm. the R.H.S. of this equation was evaluated for a range of values of $\beta$. A relationship was thus obtained between $\beta$ and Ri75 which is shown in fig. 7 using the relationships between u150/u37.5 and Ri75 and between u150/u37.5 and $\beta$. The parameter $k^2\left(\frac{z}{z_o}\right)^{2\beta-2}$ was then evaluated using k = 0.41 and zo = 0.30 cm. for a series of values of $\beta$ and finally the values of this parameter were plotted against Ri75, a plot which is shown as the broken line on fig. 8.

A plot of the values of $\tau_o \Big/ \rho z^2 \left(\frac{\partial u}{\partial z}\right)_z^2$ against the Richardson Numbers at the three heights 150, 75, and 37.5 cm. showed no separation according to height but there was a general scatter which arose from the observational inaccuracies particularly in the values of $\tau_o$ owing to the non-representative nature of the surface sample for which $\tau_o$ was measured on any one occasion. Mean values of the parameter for ranges of Richardson Number were calculated and plotted against the mean Richardson Number in each range. Fig. 8 shows these plots. Here again the number attached to each point represents the number of observations represented by the point. A smooth curve has been drawn through the points and is shown as the full line in the figure. In table VI values of the parameters corresponding to various values of the Richardson Number as read from the smooth curves of fig. 8 are set out together with other quantities, some of which will be referred to later. It is evident that there is excellent agreement

Table VI

| Ri75 x $10^3$ | -300 | -200 | -150 | -100 | -75 | -50 | -25 | 0 | +25 | +50 | +75 | +100 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $\frac{u150}{u37.5}$ | 1.193 | 1.202 | 1.206 | 1.213 | 1.222 | 1.235 | 1.254 | 1.285 | 1.321 | 1.374 | 1.445 | 1.548 |
| $\beta$ (approx) | 1.13 | 1.12 | 1.11 | 1.10 | 1.08 | 1.06 | 1.04 | 1.00 | .96 | .91 | .85 | .78 |
| $\frac{\tau_o}{\rho z^2 \left(\frac{\partial u}{\partial z}\right)_z^2}$ | .67 | .64 | .61 | .53 | .46 | .35 | .25 | .17 | .14 | .12 | .10 | .09 |
| $k^2\left(\frac{z}{z_o}\right)^{2\beta-2}$ | .70 | .64 | .58 | .49 | .42 | .33 | .25 | .17 | .11 | .07 | .03 | .01 |
| $\sigma_{RM}$ | 2.5 | 3.7 | 4.8 | 6.8 | 8.5 | 10.4 | 13.1 |  | 8.0 | 8.0 | 9.1 | 8.7 |
| $\sigma_H$ | 10.0 | 14.0 | 17.5 | 21.5 | 23.2 | 21.7 | 19.5 |  | 6.7 | 5.7 | 5.4 | 4.6 |
| $\frac{-E_o}{\left(\frac{\partial\psi}{\partial z}\right)_z\left(\frac{\partial u}{\partial z}\right)_z z^2}$ |  | .62 | .60 | .53 | .45 | .37 | .27 | .18 | .16 | .11 | .10 | .08 |

between the values of $k^2 \left(\frac{z}{z_o}\right)^{2\beta-2}$ and $\tau_o / \rho z^2 \left(\frac{\partial u}{\partial z}\right)^2_z$ in neutral and unstable conditions but that in stable conditions the values of $k^2 \left(\frac{z}{z_o}\right)^{2\beta-2}$ become progressively less than those of $\tau_o / \rho z^2 \left(\frac{\partial u}{\partial z}\right)^2_z$. The use of equation 5 therefore leads to satisfactory values of Kmz in unstable (and neutral) conditions but its use in stable conditions is not permissible and would lead to large underestimations of Km. We may now regard equation 4 as explicitly verified for unstable conditions of flow.

Deacon has suggested that zo may not be independent of stability in stable conditions and it is therefore necessary to consider the change necessary in zo to provide agreement between the values of $k^2 \left(\frac{z}{z_o}\right)^{2\beta-2}$ and the observed values of $\tau_o / \rho z^2 \left(\frac{\partial u}{\partial z}\right)^2_z$ in these conditions. To achieve this the value of zo must increase with increasing degree of stability taking, for example, a value of 1 cm. in the region of Ri75 = +50 x $10^{-3}$. It is very difficult to see any reason for such a variation in zo and it is considered that the suggestion of the dependence of zo on stability in stable conditions does not explain the breakdown of equation 4. We must also note that any likely error in the value used for k will not affect the results presented here in any material way.

Another basis on which the results may be examined is provided by the theoretical treatments due to Rossby and Montgomery (1935) and Holzman (1943). These treatments lead to the following equations

Rossby and Montgomery

$$\frac{\tau_o}{\rho z^2 \left(\frac{\partial u}{\partial z}\right)^2_z} = \frac{k^2}{(1 + \sigma \, Ri)} \qquad 7a.$$

Holzman

$$\frac{\tau_o}{\rho z^2 \left(\frac{\partial u}{\partial z}\right)^2_z} = k^2(1 - \sigma \, Ri) \qquad 7b.$$

$\sigma$ is a proportionality factor introduced in the relationship between potential and eddying energy associated with the thermal stratification. Our values of $\tau_o / \rho z^2 \left(\frac{\partial u}{\partial z}\right)^2_z$ enable values of the $\sigma$ to be calculated, and these have been included in table VI. It will be seen that both equations are unsatisfactory over the whole range of stability covered, but that the Holzman equation is better than the Rossby-Montgomery equation. Considering stable conditions only it will be seen that both these equations have reasonably constant values of $\sigma$, which has a value of about 8 in the Rossby-Montgomery equation and about 6 in the Holzman equation. Either of these equations with the appropriate value for $\sigma$ would lead to acceptable values of Km, and both appear to be superior to Deacon's equation in stable conditions. The magnitudes of $\sigma$ do not appear to have any particular significance. In table II values of Km 75 calculated from $Km75 = \tau_o / \rho \left(\frac{\partial u}{\partial z}\right)_{75}$ have been given where possible. It will be seen that the order of magnitude is in the range from $10^2$ to $10^3$.

## 5. The eddy diffusivity for vapour.

By making the assumption of a negligibly small horizontal variation of absolute humidity, we may obtain values for the eddy diffusivity for vapour at the height z, Kvz, from the gradients of absolute humidity $(\partial\psi/\partial z)$ and the rate of evaporation Eo, for we have by definition

$$Kvz = \frac{-Ez}{\left(\frac{\partial\psi}{\partial z}\right)_z} \qquad 8.$$

where Ez is the vertical flux of vapour at the height z. We also have

$$Ez - Eo = -\int_0^z \frac{\partial\psi}{\partial t}\,dz \qquad 9.$$

and it has been shown in the past e.g. Rider and Robinson (1951) and may be further demonstrated from the present results that the magnitude of the R.H.S. of equation 9 is always small compared with Eo. Even in the worst case in the present observations (No.43) Eo is about twenty times the magnitude of the $\int_0^z \frac{\partial\psi}{\partial t}\,dz$. We may therefore substitute Eo for Ez in equation 8 with negligible error and write

$$Kvz = \frac{-Eo}{\left(\frac{\partial\psi}{\partial z}\right)_z}$$

which in order to facilitate comparison with Kmz arranged in the form

$$\frac{Kmz}{z^2\left(\frac{\partial u}{\partial z}\right)_z} = \frac{\tau_o}{\rho z^2\left(\frac{\partial u}{\partial z}\right)_z^2}$$

we may write as

$$\frac{Kvz}{z^2\left(\frac{\partial u}{\partial z}\right)_z} = \frac{-Eo}{\left(\frac{\partial\psi}{\partial z}\right)_z\left(\frac{\partial u}{\partial z}\right)_z z^2} \qquad 10.$$

Values of the parameter of the R.H.S. of equation 10 are readily obtained from the results and are set out in table II. A plot of the magnitude of the parameter against the Richardson Number at the three heights 150, 75, and 37.5 cm. shows some scatter but no systematic separation of the plots according to the value of z. In order to construct a smooth curve through the points the values of the parameter have been meaned in ranges of Richardson Number and have been plotted against the mean value of the Richardson Number in each range in fig. 9. Here again the number against each point indicates the number of observations represented by the point. The smooth curve shown in fig. 9 is thought to represent the most likely variation of $-Eo/\left(\frac{\partial\psi}{\partial z}\right)_z\left(\frac{\partial u}{\partial z}\right)_z z^2$ with stability and values of this quantity corresponding to the various values of the Richardson Number have been taken from the curve and inserted in table VI. It is immediately apparent that throughout the range of stability covered by the observations the values of $-Eo/\left(\frac{\partial\psi}{\partial z}\right)_z\left(\frac{\partial u}{\partial z}\right)_z z^2$ correspond closely with those of $\tau_o/\rho z^2\left(\frac{\partial u}{\partial z}\right)_z^2$. In other words we have demonstrated on a direct observational basis the identity of the eddy diffusivities for vapour and momentum in unstable, neutral, and stable conditions which, as far as the writer is aware, has not been done previously. Pasquill (1949a) by assuming the validity of equation 2

/was

was able to show that his observed values of Kv in neutral conditions agreed with values of Km calculated on the basis of that equation. He also obtained good agreement between values of Kv and of Km when the latter were calculated on the basis of equation 4 in unstable but not in stable conditions. His failure to obtain agreement in stable conditions is now understood as it has been demonstrated in the previous section that satisfactory values of Km are not obtained by using this equation in these conditions.

## 6. The eddy diffusivity for heat.

On occasions when the sky was free from cloud in the second series of observations and measurement of the net outgoing long wave radiation was possible, measurement of the usual heat balance components was made with a view to calculating the magnitudes of the eddy diffusivity for heat, Kh. In the atmosphere near the earth's surface the balance of the heat exchange process leads to the equation

$$Rs - \left(Rlz + \lambda Eo + Qz + S + \int_o^z c_p \rho \frac{\partial T}{\partial t} dz\right) = 0 \qquad 11.$$

where Rs is the net incoming solar radiation at the surface

Rlz is the net outgoing long wave radiation at the height z

Qz is the non-radiative upward flux of heat at the height z

$\lambda$ Eo is the heat associated with evaporation where $\lambda$ is the latent heat of condensation and has been given the value of 600 cal. $g^{-1}$

S is the heat absorbed in the soil.

The magnitude of the term

$$\int_o^z c_p \rho \frac{\partial T}{\partial t} dz ,$$

which is a measure of the heat absorbed in the layer from the ground to the height z, is always small when z is not large and may be neglected in comparison with the other terms in equation 11. By measuring the other terms in the equation we may evaluate Qz and hence Khz since we have, by definition,

$$Khz = \frac{Qz}{\rho\, c_p\left(\frac{\partial T}{\partial z} + \Gamma\right)}$$

In this equation Cp, the specific heat of air at constant pressure, has been given the value 0.25 and $\rho$ , the density of air, has been taken as $1.2 \times 10^{-3}$ g cm. $^{-3}$.

There were eleven occasions on which Rlz could be estimated with confidence using Robinson's (1947) method and the magnitude of the terms in equation 11 together with the deduced value of Kh75 have been set out in table III. Values of Km75 and of Kv75 have also been added where they were available, but it must be remembered that these values are calculated on the basis of a single observation of $\tau_o$ and Eo respectively and are consequently liable to be considerably in error. Unfortunately in six of the observations values of Kv75 are not available as the humidity gradients were too small to enable values to be quoted with any confidence. Values of Km75 are available in all but two cases and we are therefore able to compare Kh75 with either or both Kv75 and Km75 in all but one observation (No. 47). It will be noted that whilst the eddy diffusivity for heat is always greater than the other two diffusivities, the difference is not large on eight occasions when the possible errors, particularly in Km75, are taken into consideration. However, in observations 35 and 48, the first a very unstable case ($Ri75 = -191 \times 10^{-3}$), the second a fairly stable case ($Ri75 = +55 \times 10^{-3}$) there is a marked difference between Kh and Km or Kv. In observation 35 Kh75 is approximately equal to three times Kv75, in observation 48 it is approximately equal to twice Km75. In neither of these cases are values of both the eddy diffusivities for momentum and vapour available to enable a check of one to be made on the other, but there is no reason to suppose that the values given are grossly in error. It is equally impossible to see how any error in the heat balance components sufficiently large to reduce the magnitude of Kh75 to that of the other diffusivity could have occurred on either occasion. It is noteworthy that in observation 48 we have a case in which the humidity profile appeared to have a functional form different from that of the temperature and wind profiles, but that in observation 35 the profiles of the elements seemed to have the same form. It has been pointed out by Rider and Robinson (1951) that it is very difficult to understand why the diffusivities for heat, vapour and momentum should not be identical when the profiles have the same form as this implies that the ratios of the appropriate gradients, and hence of the diffusivities themselves, must be constant with height. The observations presented here do not support Pasquill's result that there is approximate equality between the diffusivities for heat and vapour in stable conditions and an increasing divergence between them as one passes to unstable conditions, Kh becoming progressively greater than Kv. They indicate that for the majority of occasions, irrespective of stability, there is approximately equality between the diffusivity for heat and the other two diffusivities but that exceptions occur when former diffusivity becomes larger than either of the other two. No indication of the proportion of occasions on which this occurs can be given at present, and further investigation on this matter is called for.

## 7. Conclusions.

The observational study presented here leads to the following conclusions:

(a) The temperature and wind profiles up to 2m. invariably have the same form and this form is shared by the humidity profile on the majority of occasions.

(b) In neutral conditions the eddy diffusivity for momentum may be calculated from wind profile measurements using the equation

$$Kmz = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)_z$$

In unstable conditions the eddy diffusivity for momentum is adequately specified on the basis of Deacon's wind profile law using the equation

$$Kmz = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)_z \left(\frac{z}{z_o}\right)^{2\beta-2}$$

In stable conditions the use of this equation leads to values of the diffusivity which are too small.
In stable conditions both the Holzman and the Rossby-Montgomery equations

$$Kmz = k^2 z^2 \frac{(\partial u)}{(\partial z)_z} (1 - \sigma Ri)$$

$$Kmz = \frac{k^2 z^2 \frac{(\partial u)}{(\partial z)_z}}{(1 + \cdot \sigma Ri)}$$

provide acceptable values of the eddy diffusivity for momentum with $\sigma$ equal to 6 in the former and 8 in the latter. Neither equation holds in unstable conditions with a single value of $\sigma$.

(c) The magnitude of the eddy diffusivity for momentum is identical with that for vapour irrespective of stability i.e.

$$Kmz \approx Kvz.$$

(d) The magnitude of the eddy diffusivity for heat is often the same as that for momentum or vapour but exceptions occur when it is larger than the other two. The exceptions do not appear to be related to stability.

/References

| | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 179 | 144 | 123 | 62 | 48 | 65 | 48 | 225 | 214 | 212 | 162 | 102 | 120 | 133 | 159 |
| -14 | - | 13 | 13 | 25 | 28 | 27 | 14 | 16 | 21 | 16 | 25 | 16 | 19 | 16 |
| -32 | -8 | 26 | 28 | 43 | 43 | 57 | 37 | 32 | 36 | 47 | 51 | 29 | 47 | 40 |
| -80 | -15 | 55 | 47 | 82 | 74 | 139 | 64 | 53 | 59 | 106 | 90 | 67 | 78 | 75 |
| - | -15 | -12 | -12 | - | - | - | - | - | - | - | - | - | - | - |
| -32 | -30 | -23 | -25 | - | - | 11 | -10 | - | -8 | - | - | - | - | - |
| -122 | -56 | -50 | -42 | 9 | 11 | 26 | -17 | -7 | -13 | -27 | -17 | - | - | -12 |
| -54 | - | 46 | 128 | 373 | 152 | - | 18 | 14 | 13 | 84 | 108 | 63 | 59 | 49 |
| -22 | -5 | 26 | 107 | 235 | 108 | 546 | 7 | 7 | 7 | 32 | 55 | 35 | 25 | 19 |
| -9 | -3 | 12 | 41 | 123 | 61 | 211 | 4 | 4 | 4 | 14 | 30 | 16 | 16 | 10 |
| .762 | .520 | .308 | .080 | .026 | .085 | .028 | 1.54 | 2.23 | 2.59 | - | .191 | .244 | .457 | .403 |
| 30 | 14 | 24 | 9 | 4 | 4 | - | 23 | 21 | 12 | - | 9 | 11 | 15 | 13 |
| 22 | 12 | 14 | 9 | 6 | 8 | 12 | 14 | 26 | 23 | - | 9 | 13 | 10 | 8 |
| 14 | 15 | 6 | 10 | 7 | 12 | 7 | 18 | 29 | 34 | - | 11 | 10 | 15 | 9 |
| 355 | 245 | 105 | 38 | - | - | -16 | 165 | 24 | 165 | 60 | 43 | -75 | -26 | 24 |
| - | 20 | 15 | 18 | - | - | - | - | - | - | - | - | - | - | - |
| 28 | 19 | 15 | 7 | - | - | 14 | 23 | - | 11 | - | - | - | - | - |
| 12 | 28 | 15 | 10 | - | - | 9 | 31 | 12 | 17 | 12 | 15 | - | - | 9 |
| 1110 | 820 | 460 | 150 | - | - | 150 | 1650 | - | 2060 | - | - | - | - | - |
| 900 | 560 | 450 | 180 | 90 | 190 | 120 | 990 | 1740 | 1670 | - | 280 | 370 | 470 | 390 |

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## 7. Conclusions.

The observational study presented here leads to the following conclusions:

(a) The temperature and wind profiles up to 2m. invariably have the same form and this form is shared by the humidity profile on the majority of occasions.

(b) In neutral conditions the eddy diffusivity for momentum may be calculated from wind profile measurements using the equation

$$K_{mz} = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)_z$$

In unstable conditions the eddy diffusivity for momentum is adequately specified on the basis of Deacon's wind profile law using the equation

$$K_{mz} = k^2 z^2 \left(\frac{\partial u}{\partial z}\right)_z \left(\frac{z}{z_0}\right)^{2\beta-2}$$

/In

BLE II

| 37 | 38 | 39 | 40 | 41 | 42 | 43 | 44 | 45 | 46 | 47 | 48 | 49 | 50 | 51 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 17/10 | 17/10 | 17/10 | 17/10 | 17/10 | 17/10 | 17/10 | 24/10 | 24/10 | 24/10 | 7/11 | 7/11 | 7/11 | 7/11 | 7/11 |
| 1415 | 1530 | 1655 | 1830 | 1930 | 2030 | 2130 | 1815 | 2000 | 2100 | 1715 | 1815 | 1915 | 2015 | 2115 |
| 369 | 339 | 254 | 126 | 70 | 108 | 66 | 577 | 591 | 657 | 326 | 264 | 249 | 307 | 342 |
| 352 | 326 | 236 | 115 | 64 | 96 | 62 | 549 | 563 | 624 | 310 | 250 | 239 | 292 | 328 |
| 330 | 302 | 216 | 101 | 51 | 81 | 54 | 506 | 518 | 575 | 291 | 232 | 222 | 272 | 305 |
| 316 | 284 | 208 | 99 | 50 | 74 | 49 | 490 | 511 | 562 | 273 | 214 | 209 | 253 | 284 |
| 288 | 263 | 186 | 81 | 41 | 61 | 44 | 442 | 453 | 504 | 252 | 197 | 193 | 232 | 262 |
| 273 | 250 | 180 | 78 | 37 | 56 | 41 | 427 | 438 | 490 | 239 | 187 | 179 | 221 | 251 |
| 254 | 231 | 165 | 71 | 33 | 50 | 35 | 390 | 407 | 462 | 216 | 168 | 160 | 197 | 223 |
| 223 | 203 | 144 | 59 | - | 41 | 30 | 352 | 364 | 428 | 190 | 146 | 142 | 172 | 166 |
| 55.2 | 55.6 | 54.3 | 53.2 | 45.4 | 43.9 | 43.3 | 47.3 | 47.4 | 48.1 | 42.5 | 40.1 | 39.0 | 39.5 | 40.2 |
| 55.26 | 55.60 | 54.19 | 53.07 | 45.24 | 43.66 | 43.10 | 47.18 | 47.29 | 47.98 | 42.41 | 39.94 | 38.88 | 39.39 | 40.09 |
| 55.47 | 55.64 | 54.05 | 52.91 | 44.99 | 43.43 | 42.82 | 47.11 | 47.10 | 47.76 | 42.18 | 39.66 | 38.69 | 39.23 | 39.92 |
| 55.71 | 55.65 | 53.81 | 52.67 | 44.60 | 43.00 | 42.29 | 46.71 | 46.82 | 47.42 | 41.76 | 39.18 | 38.43 | 38.78 | 39.59 |
| 55.93 | 55.78 | 53.73 | 52.58 | 44.41 | 42.87 | 41.94 | 46.60 | 46.72 | 47.32 | 41.53 | 39.00 | 38.28 | 38.64 | 39.38 |
| 56.21 | 55.84 | 53.55 | 52.43 | 44.07 | 42.60 | 41.32 | 46.40 | 46.49 | 47.07 | 41.18 | 38.69 | 38.10 | 38.37 | 39.23 |
| 6.94 | 6.77 | 6.81 | 7.04 | 6.46 | 6.32 | 6.19 | 6.87 | 6.89 | 6.75 | 4.91 | 5.02 | 6.27 | 5.31 | 5.15 |
| 6.95 | 6.78 | 6.81 | 7.09 | 6.44 | 6.28 | 6.18 | 6.89 | 6.89 | 6.76 | 4.91 | 5.02 | 6.28 | 5.29 | 5.15 |
| 6.98 | 6.82 | 6.96 | 7.16 | 6.43 | 6.26 | 6.15 | 6.88 | 6.90 | 6.78 | 4.91 | 5.02 | 6.30 | 5.31 | 5.16 |
| 7.14 | 6.95 | 6.97 | 7.28 | 6.40 | 6.24 | 6.09 | 6.95 | 6.92 | 6.83 | 4.92 | 5.03 | 6.31 | 5.30 | 5.19 |
| 7.33 | 7.07 | 7.05 | 7.34 | 6.39 | 6.21 | 6.07 | 6.95 | 6.93 | 6.85 | 4.96 | 5.05 | 6.32 | 5.30 | 5.21 |
| 7.49 | 7.34 | 7.11 | 7.41 | 6.38 | 6.21 | 6.00 | 7.00 | 6.98 | 6.88 | 4.99 | 5.07 | 6.33 | 5.30 | 5.27 |
| 31 | 37 | 30 | 18 | 15 | 25 | - | 50 | 62 | 74 | 25 | 28 | 29 | 33 | 33 |
| 71 | 78 | 57 | 37 | 25 | 37 | 19 | 130 | 128 | 130 | 71 | 57 | 53 | 81 | 85 |
| 179 | 144 | 123 | 62 | 48 | 65 | 48 | 225 | 214 | 212 | 162 | 102 | 120 | 133 | 159 |
| -14 | - | 13 | 13 | 25 | 28 | 27 | 14 | 16 | 21 | 16 | 25 | 16 | 19 | 16 |
| -32 | -8 | 26 | 28 | 43 | 43 | 57 | 37 | 32 | 36 | 47 | 51 | 29 | 47 | 40 |
| -80 | -15 | 55 | 47 | 82 | 74 | 139 | 64 | 53 | 59 | 106 | 90 | 67 | 78 | 75 |
| - | -15 | -12 | -12 | - | - | - | - | - | - | - | - | - | - | - |
| -32 | -30 | -23 | -25 | - | - | 11 | -10 | - | -8 | - | - | - | - | - |
| -122 | -56 | -50 | -42 | 9 | 11 | 26 | -17 | -7 | -13 | -27 | -17 | - | - | -12 |
| -54 | - | 46 | 128 | 373 | 152 | - | 18 | 14 | 13 | 84 | 108 | 63 | 59 | 49 |
| -22 | -5 | 26 | 107 | 235 | 108 | 546 | 7 | 7 | 7 | 32 | 55 | 35 | 25 | 19 |
| -9 | -3 | 12 | 41 | 123 | 61 | 211 | 4 | 4 | 4 | 14 | 30 | 16 | 16 | 10 |
| .762 | .520 | .308 | .080 | .026 | .085 | .028 | 1.54 | 2.23 | 2.59 | - | .191 | .244 | .457 | .403 |
| 30 | 14 | 24 | 9 | 4 | 4 | - | 23 | 21 | 12 | - | 9 | 11 | 15 | 13 |
| 22 | 12 | 14 | 9 | 6 | 8 | 12 | 14 | 26 | 23 | - | 9 | 13 | 10 | 8 |
| 14 | 15 | 6 | 10 | 7 | 12 | 7 | 18 | 29 | 34 | - | 11 | 10 | 15 | 9 |
| 355 | 245 | 105 | 38 | - | - | -16 | 165 | 24 | 165 | 60 | 43 | -75 | -26 | 24 |
| - | 20 | 15 | 18 | - | - | - | - | - | - | - | - | - | - | - |
| 28 | 19 | 15 | 7 | - | - | 14 | 23 | - | 11 | - | - | - | - | - |
| 12 | 28 | 15 | 10 | - | - | 9 | 31 | 12 | 17 | 12 | 15 | - | - | 9 |
| 1110 | 820 | 460 | 150 | - | - | 150 | 1650 | - | 2060 | - | - | - | - | - |
| 900 | 560 | 450 | 180 | 90 | 190 | 120 | 990 | 1740 | 1670 | - | 280 | 370 | 470 | 390 |

T A

| | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Obs. No. | | 27 | 28 | 29 | 30 | 31 | 32 | 33 | 34 | 35 | 36 |
| Date | (1952) | 8/10 | 8/10 | 8/10 | 8/10 | 10/10 | 10/10 | 15/10 | 15/10 | 16/10 | 16/1 |
| Time | G.M.T. | 1445 | 1645 | 1915 | 2145 | 1100 | 1315 | 1410 | 1500 | 1118 | 1400 |
| Wind speed u cm. sec.$^{-1}$ at z cm. | 200 | 437 | 608 | 238 | 393 | 374 | 368 | 215 | 235 | 157 | 199 |
| | 150 | 409 | 571 | 216 | 368 | 360 | 355 | 205 | 223 | 152 | 192 |
| | 100 | 379 | 525 | 194 | 333 | 341 | 337 | 194 | 207 | 143 | 181 |
| | 75 | 362 | 508 | 182 | 325 | 327 | 327 | 186 | 194 | 135 | 173 |
| | 50 | 336 | 463 | 164 | 292 | 307 | 302 | 173 | 184 | 130 | 159 |
| | $37\frac{1}{2}$ | 320 | 449 | 154 | 279 | 292 | 286 | 166 | 174 | 125 | 150 |
| | 25 | 272 | 405 | 140 | 260 | 267 | 262 | 148 | 157 | 117 | 141 |
| | 15 | 250 | 352 | 121 | 225 | 237 | 230 | 134 | 136 | 102 | 124 |
| Air temperature T°F at z cm. | 200 | 54.0 | 54.1 | 46.3 | 46.0 | 49.1 | 51.3 | 51.0 | 51.6 | 47.7 | 51.6 |
| | 150 | 54.01 | 54.01 | 45.98 | 45.84 | 49.25 | 51.36 | 51.03 | 51.62 | 47.79 | 51.6 |
| | 100 | 54.01 | 53.91 | 45.64 | 45.63 | 49.48 | 51.52 | 51.15 | 51.63 | 48.05 | 51.6 |
| | 50 | 54.03 | 53.74 | 45.18 | 45.26 | 50.01 | 51.83 | 51.37 | 51.63 | 48.38 | 51.7 |
| | $37\frac{1}{2}$ | 54.03 | 53.68 | 45.02 | 45.13 | 50.26 | 51.93 | 51.62 | 51.71 | 48.66 | 51.7 |
| | 25 | 54.07 | 53.60 | 44.77 | 44.93 | 50.64 | 52.18 | 51.80 | 51.73 | 48.92 | 51.8 |
| Absolute humidity $\psi$ g. m.$^{-3}$ | 200 | 7.34 | 7.31 | 6.91 | 6.62 | 5.64 | 5.44 | 6.07 | 6.40 | 6.48 | 5.68 |
| | 150 | 7.36 | 7.33 | 6.89 | 6.62 | 5.68 | 5.47 | 6.03 | 6.47 | 6.50 | 5.69 |
| | 100 | 7.38 | 7.36 | 6.84 | 6.63 | 5.81 | 5.49 | 6.18 | 6.56 | 6.49 | 5.74 |
| | 50 | 7.51 | 7.45 | 6.82 | 6.65 | 6.00 | 5.65 | 6.33 | 6.77 | 6.66 | 5.90 |
| | $37\frac{1}{2}$ | 7.60 | 7.51 | 6.82 | 6.67 | 6.11 | 5.75 | 6.36 | 6.81 | 6.77 | 6.02 |
| | 25 | 7.73 | 7.63 | 6.80 | 6.69 | 6.32 | 5.94 | 6.64 | 6.85 | 6.90 | 6.19 |
| $\frac{\partial u}{\partial z} \times 10^2$ sec.$^{-1}$ | 150 | 53 | 71 | 44 | 51 | 31 | 26 | 17 | 25 | 13 | 14 |
| | 75 | 83 | 134 | 62 | 87 | 64 | 60 | 39 | 44 | 27 | 41 |
| | $37\frac{1}{2}$ | 202 | 232 | 88 | 151 | 157 | 171 | 102 | 99 | 65 | 97 |
| $\frac{\partial T}{\partial z} \times 10^4$ °C cm.$^{-1}$ | 150 | - | 10 | 37 | 23 | -26 | -12 | -11 | - | -18 | - |
| | 75 | - | 19 | 52 | 39 | -53 | -28 | -26 | - | -39 | -9 |
| | $37\frac{1}{2}$ | -4 | 33 | 74 | 69 | -130 | -81 | -67 | -8 | -93 | -22 |
| $\frac{\partial \psi}{\partial z} \times 10^{10}$ g. cm.$^{-4}$ | 150 | - | -10 | - | - | -19 | -11 | -14 | -16 | -11 | -11 |
| | 75 | -26 | -20 | - | - | -39 | -25 | -32 | -27 | -25 | -32 |
| | $37\frac{1}{2}$ | -82 | -34 | 10 | -15 | -95 | -72 | -82 | -61 | -58 | -76 |
| Ri x $10^3$ | 150 | - | 6 | 65 | 30 | -98 | -66 | -144 | - | -392 | - |
| | 75 | - | 3 | 46 | 14 | -45 | -29 | -61 | - | -191 | -21 |
| | $37\frac{1}{2}$ | 0 | 2 | 33 | 10 | -18 | -10 | -23 | -3 | -77 | -11 |
| $\tau_o$ dynes cm.$^{-2}$ | | 1.22 | 1.11 | .296 | .982 | 1.04 | .850 | .342 | .373 | - | - |
| $\frac{\tau_o}{\rho z^2 \left(\frac{\partial u}{\partial z}\right)^2} \times 10^2$ | 150 | 16 | 8 | 6 | 13 | 41 | 47 | 44 | 22 | - | - |
| | 75 | 26 | 10 | 11 | 19 | 37 | 35 | 33 | 29 | - | - |
| | $37\frac{1}{2}$ | 18 | 13 | 22 | 26 | 25 | 17 | 20 | 22 | - | - |
| Eo x $10^8$ g. cm.$^{-2}$ sec.$^{-1}$ | | 348 | 205 | -28 | 51 | 567 | 322 | 315 | 257 | 233 | 200 |
| $\frac{-\,Eo}{z^2 \left(\frac{\partial u}{\partial z}\right)\left(\frac{\partial \psi}{\partial z}\right)}$ | 150 | - | 13 | - | - | 43 | 50 | 60 | 29 | 72 | 58 |
| | 75 | 29 | 14 | - | - | 40 | 38 | 45 | 38 | 61 | 27 |
| | $37\frac{1}{2}$ | 15 | 19 | 22 | 15 | 27 | 19 | 27 | 30 | 44 | 19 |
| Kv75 cm.$^2$ sec.$^{-1}$ | | 1340 | 1020 | - | - | 1450 | 1290 | 980 | 950 | 930 | 620 |
| Km75 cm.$^2$ sec.$^{-1}$ | | 1230 | 690 | 400 | 940 | 1360 | 1180 | 730 | 710 | - | - |

# TABLE III

| Observation No. | 29 | 30 | 31 | 33 | 35 | 44 | 45 | 46 | 47 | 48 | 50 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Heat Components in Cal. cm.$^{-2}$ min.$^{-1}$ | | | | | | | | | | | |
| Net Incoming Solar Radiation | - | - | .568 | .329 | .459 | - | - | - | - | - | - |
| Net Outward Long Wave Radiation at 75 cm. | .104 | .102 | .168 | .142 | .144 | .122 | .122 | .118 | .119 | .116 | .115 |
| Heat Gained by Soil | -.043 | -.024 | .046 | .015 | .043 | -.055 | -.024 | -.035 | -.030 | -.082 | -.056 |
| Heat Used in Evaporation | -.010 | .018 | .204 | .114 | .084 | .059 | .009 | .059 | .022 | .015 | -.009 |
| Heat Associated With Turbulent Transport | -.051 | -.096 | .150 | .058 | .188 | -.126 | -.107 | -.142 | -.111 | -.049 | -.050 |
| Kh75 cm.$^2$ sec.$^{-1}$ | 560 | 1400 | 1540 | 1190 | 2610 | 1940 | 1920 | 2250 | 1340 | 590 | 600 |
| Kv75 cm.$^2$ sec.$^{-1}$ (From Table II) | - | - | 1450 | 980 | 930 | 1650 | - | 2060 | - | - | - |
| Km75 cm.$^2$ sec.$^{-1}$ (From Table II) | 400 | 940 | 1360 | 730 | - | 990 | 1740 | 1670 | - | 280 | 470 |

# T A B L E V

Observations of wind speed profile and aerodynamic drag in near neutral conditions as specified by $|Ri_{75}| \leqslant 10 \times 10^{-3}$.

| Obs. Number | $Ri_{75}$ x $10^3$ | $u_{75}$ cm. $sec.^{-1}$ | $u_z/u_{75}$ at z cm. = 200 | 150 | 100 | 75 | 50 | $37\frac{1}{2}$ | 25 | 15 | $\tau_o$ dynes $cm.^{-2}$ | $\frac{u_{75}}{\tau_o^{1/2}}$ |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | -10 | 521 | 1.140 | 1.106 | 1.043 | 1.000 | .919 | .858 | .762 | .637 | 2.35 | 341 |
| 3 | -10 | 585 | 1.125 | 1.084 | 1.029 | | .928 | .867 | .756 | .658 | 2.97 | 338 |
| 10 | -9 | 331 | 1.169 | 1.116 | 1.077 | | .925 | .876 | .787 | .707 | .607 | 424 |
| 11 | -4 | 669 | 1.196 | 1.145 | 1.093 | | .909 | .839 | .742 | .613 | 3.73 | 347 |
| 12 | -5 | 749 | 1.139 | 1.108 | 1.070 | | .918 | .853 | .749 | .640 | 3.81 | 384 |
| 13 | -4 | 822 | 1.114 | 1.095 | 1.040 | | .926 | .866 | .753 | .627 | 3.04 | 472 |
| 15 | -5 | 546 | 1.149 | 1.097 | 1.055 | | .919 | .866 | .771 | .696 | 1.96 | 390 |
| 19 | -10 | 381 | 1.150 | 1.100 | 1.055 | | .908 | .873 | .750 | .643 | .809 | 423 |
| 22 | -3 | 463 | 1.157 | 1.104 | 1.041 | | .910 | .875 | .771 | .652 | 1.63 | 362 |
| 28 | +3 | 508 | 1.197 | 1.124 | 1.034 | | .911 | .884 | .797 | .693 | 1.11 | 484 |
| 38 | -5 | 284 | 1.194 | 1.148 | 1.064 | | .919 | .880 | .838 | .715 | .520 | 394 |
| 44 | +7 | 490 | 1.178 | 1.120 | 1.033 | | .903 | .872 | .796 | .718 | 1.54 | 395 |
| 45 | +7 | 511 | 1.156 | 1.101 | 1.014 | | .886 | .857 | .796 | .713 | 2.23 | 343 |
| 46 | +7 | 562 | 1.169 | 1.110 | 1.023 | | .897 | .872 | .822 | .762 | 2.59 | 287 |
| Means | -3 | | 1.160 | 1.111 | 1.048 | 1.000 | .913 | .868 | .778 | .678 | | 385 |

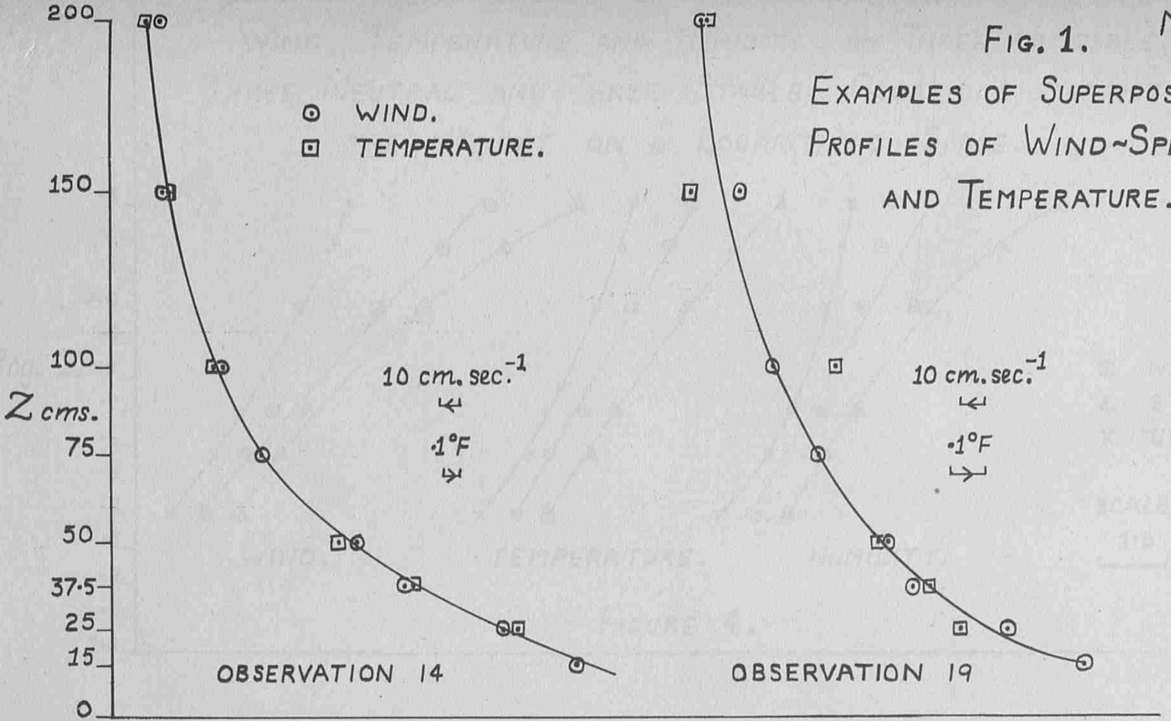

FIG. 1.
EXAMPLES OF SUPERPOSED PROFILES OF WIND-SPEED AND TEMPERATURE.

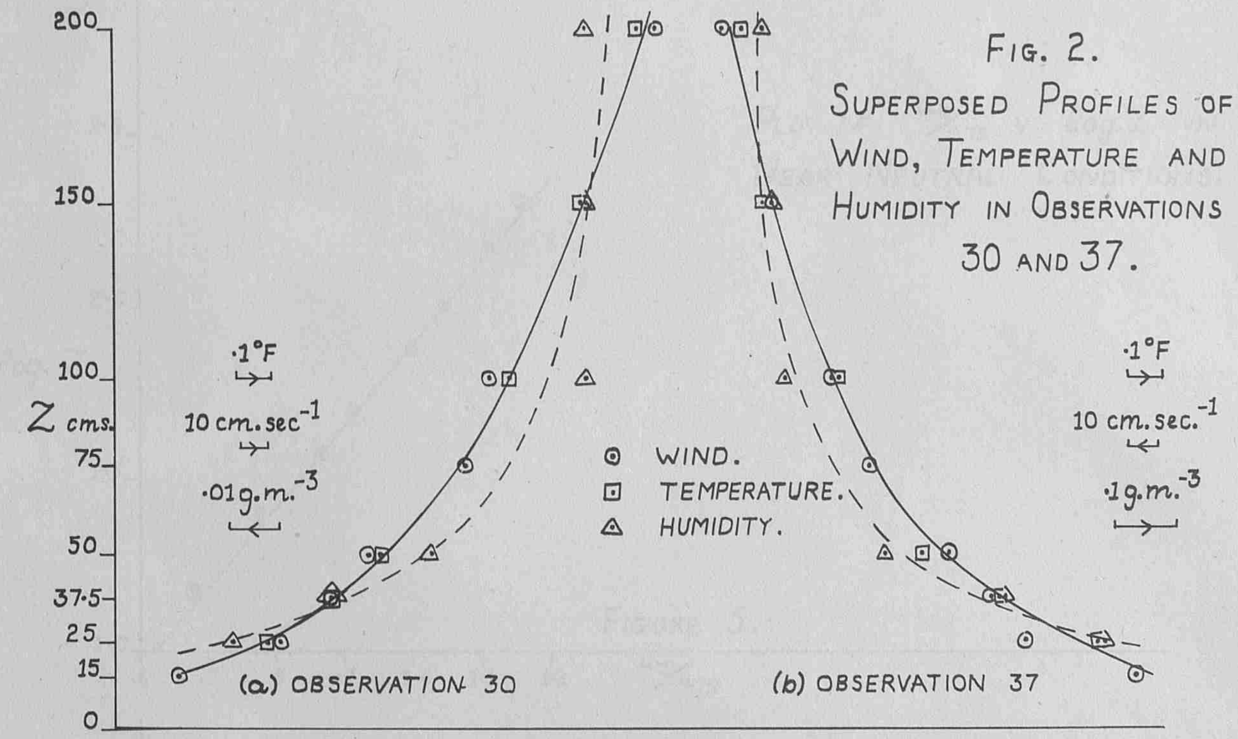

FIG. 2.
SUPERPOSED PROFILES OF WIND, TEMPERATURE AND HUMIDITY IN OBSERVATIONS 30 AND 37.

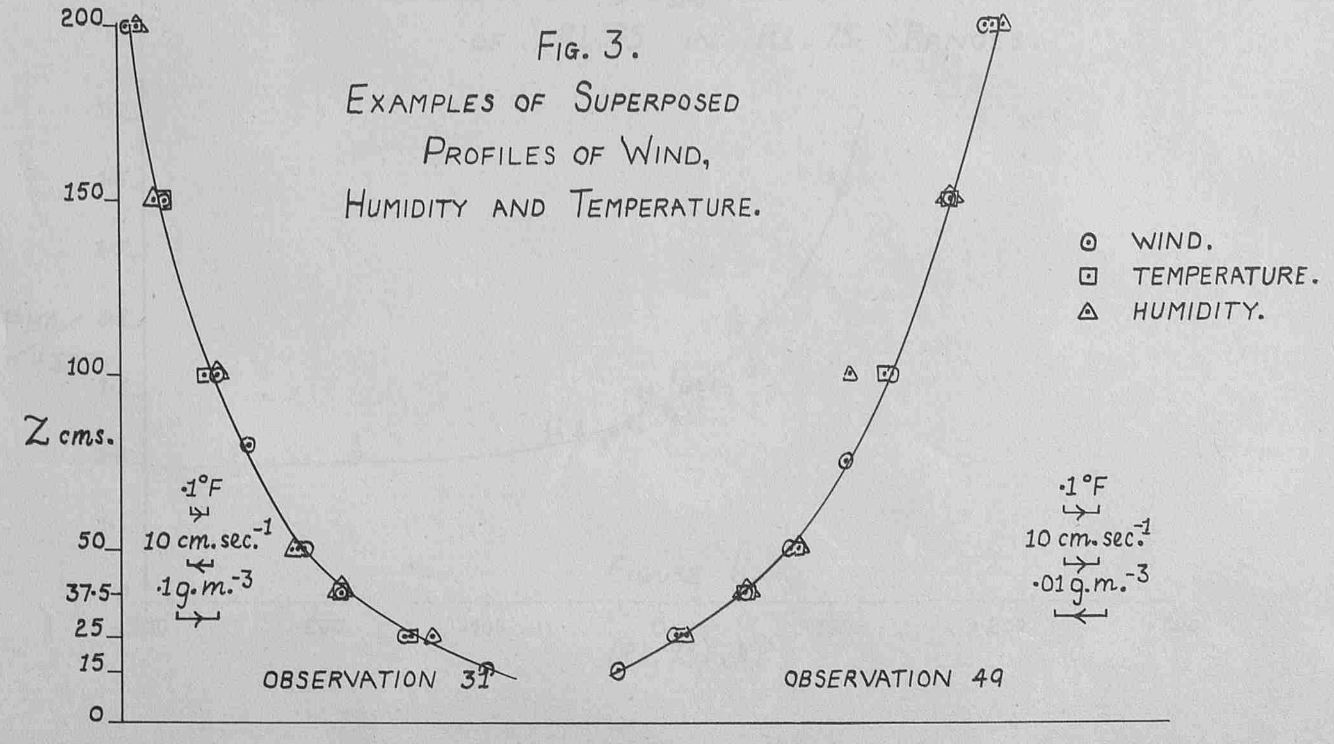

FIG. 3.
EXAMPLES OF SUPERPOSED PROFILES OF WIND, HUMIDITY AND TEMPERATURE.

Plots of Mean Values of the Parameter $\frac{S_{25}-S_z}{S_{25}-S_{50}}$ for Wind, Temperature and Humidity on Three Unstable, Three Neutral and Three Stable Occasions Against the Height on a Logarithmic Scale.

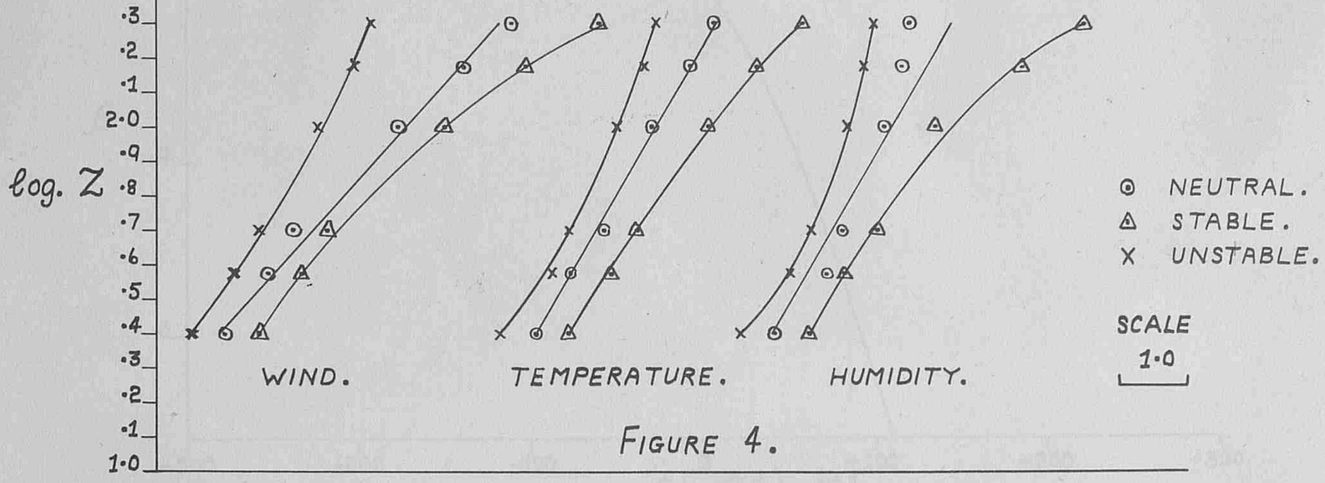

Figure 4.

Plot of $u_z/u_{75}$ v log. Z in Near Neutral Conditions.

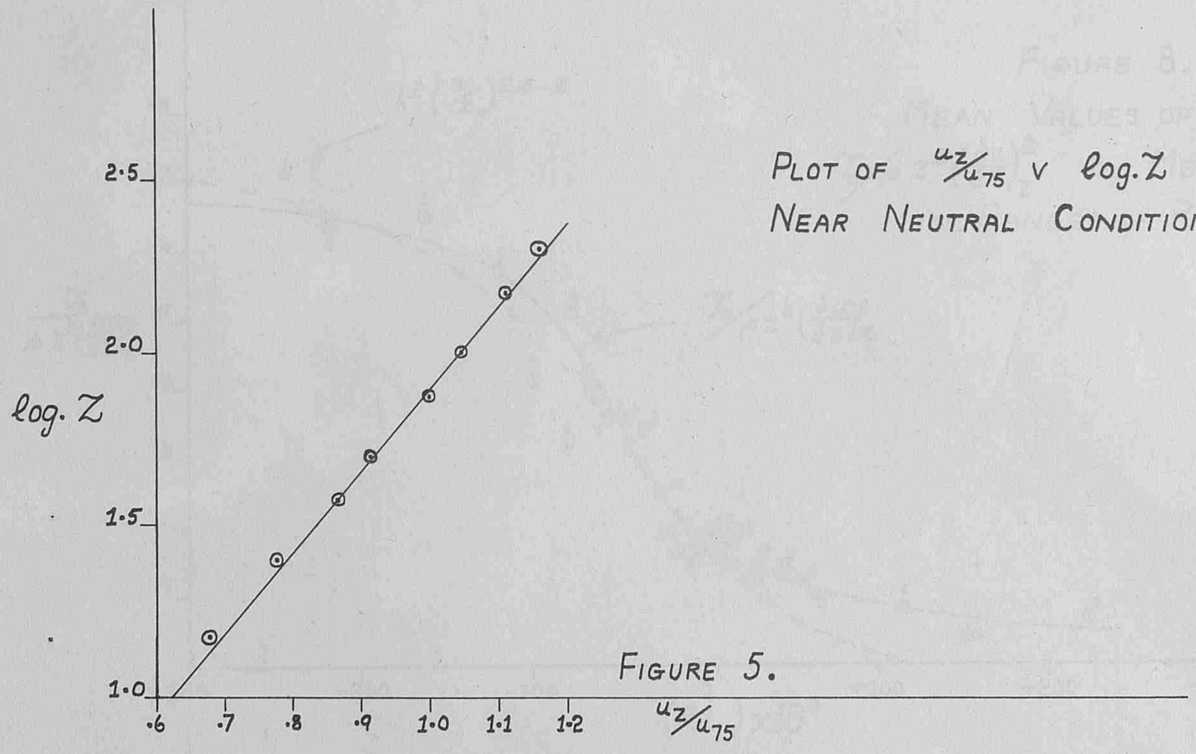

Figure 5.

Mean Values of $u_{150}/u_{37·5}$ Plotted against Mean Values of Ri. 75 in Ri. 75 Ranges.

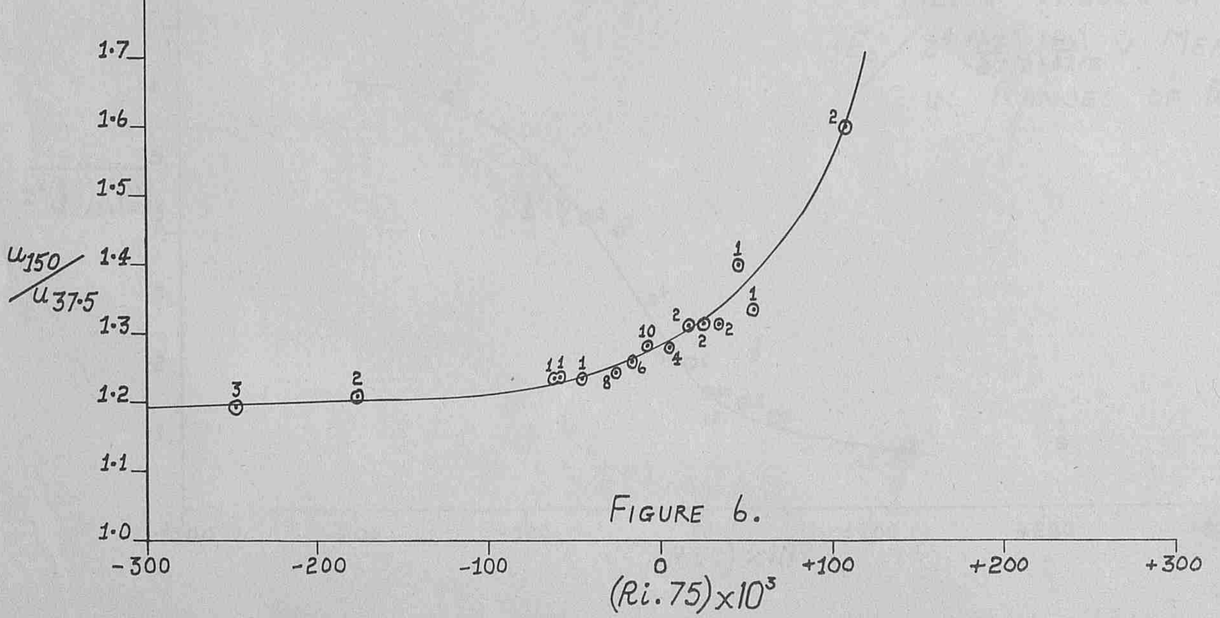

Figure 6.

M R P 809

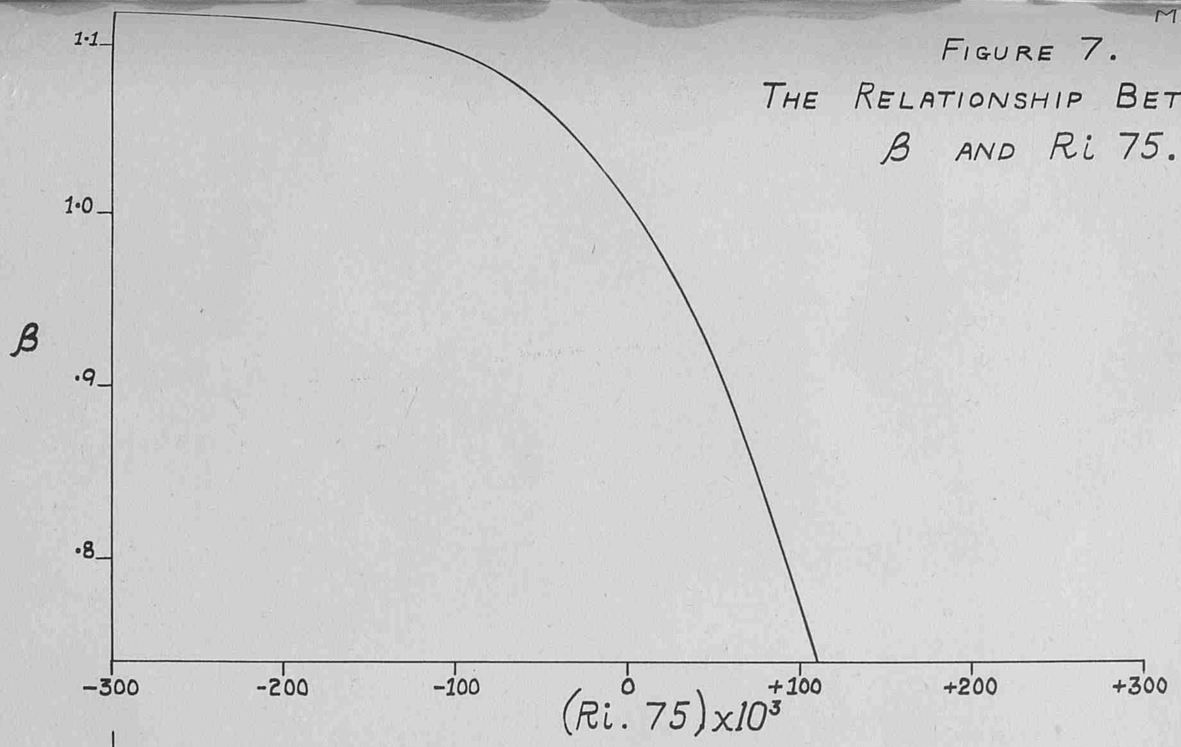

FIGURE 7.
THE RELATIONSHIP BETWEEN
$\beta$ AND Ri 75.

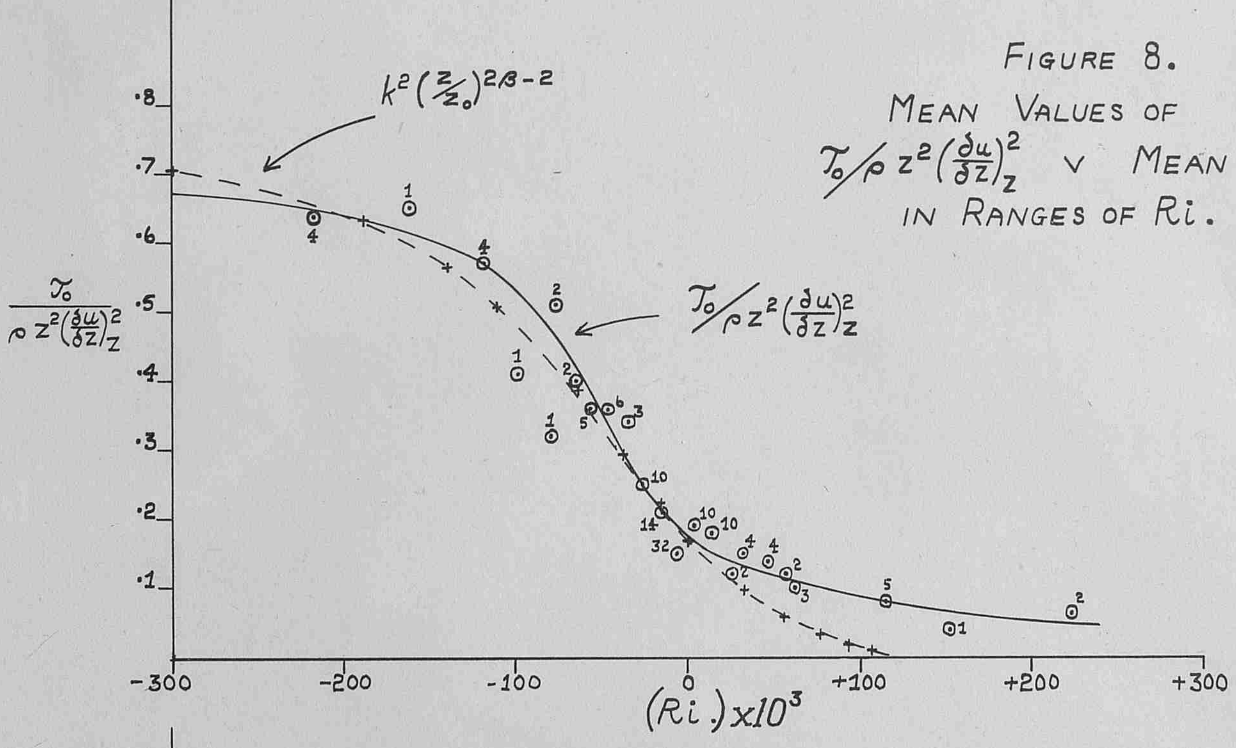

FIGURE 8.
MEAN VALUES OF
$\tau_0/\rho z^2(\frac{\partial u}{\partial z})^2_z$ V MEAN Ri
IN RANGES OF Ri.

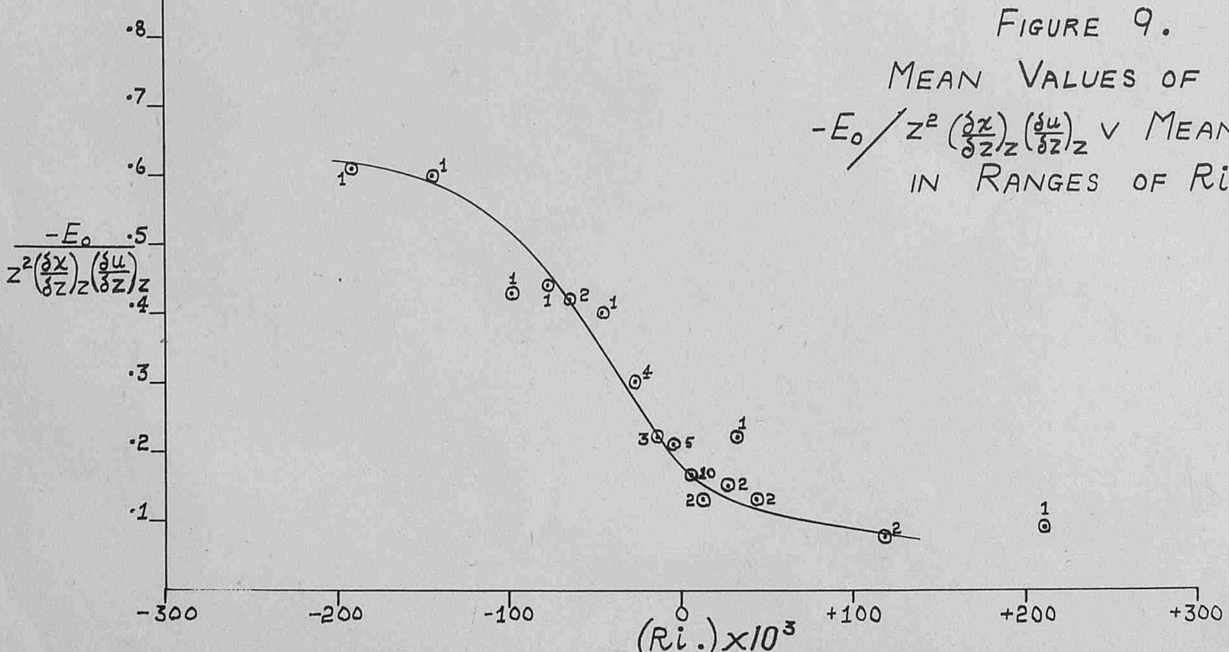

FIGURE 9.
MEAN VALUES OF
$-E_0/z^2(\frac{\partial x}{\partial z})_z(\frac{\partial u}{\partial z})_z$ V MEAN Ri
IN RANGES OF Ri.

MRP's 810, 811, 812, 813, 816

Mats

At the time of issue this report has not
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M.R.P. 810

S.C.III/153

14th May, 1953.

AD 33155

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# The use of aircraft to measure wind shear by observation of vertical smoke trails

by

I.J.W. Pothecary
R.J. Murgatroyd (Meteorological Office)

### Summary.

Vertical smoke trails have been produced by dropping smoke generators from an aircraft in flight. By observing and photographing the subsequent development of the trail from a second aircraft it is possible to measure the vertical wind shear.

### Introduction.

In M.R.P. 699 a description is given of the performance of six smoke generators made by C.D.E.E. Porton for the study of turbulence at high levels. Horizontal trails about 5 miles long were produced from continuously emitting generators fastened under the wing of a Mosquito aircraft up to a height of 35,000 feet.

Following this work C.D.E.E. were requested to design a generator from which a vertical smoke trail could be produced. The method to be used in this case was to ignite the generator and then drop it from the aircraft. It was anticipated that the descent of the generator would quickly become vertical and that it would produce a vertical smoke trail as it fell. The time of burning was adjusted so that the trail produced would be about 5,000 feet long.

The deformation of both horizontal and vertical smoke trails in the atmosphere will give information on wind shears and turbulence. Hence provided that accurate measurements can be made of this deformation these smoke trails should prove to be useful tools in investigating the nature of clear air turbulence. With a well designed ground installation these measurements can be readily obtained in clear weather. Unfortunately however clear air turbulence at high levels is not a very frequent phenomenon and the likelihood of it occurring in clear weather conditions over a ground installation is small. Attempts have therefore been made to make measurements of the motion of these trails from a second aircraft. The difficulties of making angular measurements of sufficient accuracy from an aircraft are considerable and uncertainty of the aircraft's and the trail's position in space further complicates the problem. It does however appear to be practicable to measure vertical wind shear to a reasonable accuracy and the method used is described in detail below.

### The vertical smoke trail generators.

(a) These generators were essentially the same as those described in M.R.P. 699 and were ignited in the same way. The outer metal case however

DSS
T/F

/was

(W1-V1 [illegible])

was replaced by a canvas bag designed to burn away completely by the end of the burning time so that nothing should reach the ground. As however the ballistic performance of this cylindrical canvas bag burning away as it fell was doubtful it was decided to do the experimental work well over the sea rather than a bombing range; otherwise it might have missed the range altogether when dropped from 30,000 feet or more if it failed to ignite. Out of the six prototypes supplied five worked successfully and one did not ignite due to the non-operation of the micro-switch in the release mechanism used.

(b) The fuse of these generators operated almost instantaneously. They were mounted in a light bomb carrier at the end of the wing of a Mosquito aircraft and it was considered advisable that they should be well clear of the aircraft before ignition actually took place. After consultation with a representative of Bomb Group R.A.E. it was arranged that the firing circuit should be completed through a micro-switch. This switch was not made until a cord about three feet long attached to the generator became taut and removed a pin as the generator fell away. Small plugs attached to the firing leads which were slightly longer than the cord were also pulled out by the generator. On the occasion when there was a failure it was found that the micro-switch pin had not pulled out and the cord had broken. The tightness of fit of this pin was rather critical. It it had been too loose there would have been a danger of it falling out due to vibration; if too tight the snatch of the cord as the generator fell was liable to break the cord rather than remove the pin.

(c) As neither the firing system nor the ballistic performance of the generator is satisfactory, recommendations for their improvement are suggested in the last section of this report.

(d) In all the experiments a second aircraft, a Hastings, was used to observe the trail. Details of the flight plan and methods of observation are given below. A summary of the results of the six experiments is given in Table 1.

(e) Figs. 1 and 2 show examples of the photographs of the development of the trail in Experiment No. 4 and the photographs in Fig. 3 were taken in Experiment No. 6.

Experiments 1 to 3 demonstrated that the rate of smoke emission of these generators is adequate up to the release height of 20,000 feet, but that the trails become too long and thin and disperse too rapidly when released at greater heights. Experiment No. 4 was the most successful and is described in detail below.

## The method of measurement.

When an aircraft is used as the platform from which measurements are made the only information readily available is the relative motion between different parts of the trail. This can be translated into a measurement of vertical wind shear provided that during the period of observation the wind at any level does not vary with time. In addition the apparent peaks and troughs of the deformation of the trail must be then dealt with as though they remain at the same height throughout the observations.

It is not possible to use the top or bottom points of the trail as reference points as dissipation at the ends usually occurs more quickly than at the centre of the trail. Hence peaks and troughs must be used as reference points. This is a weakness inherent in using the continuous line of a trail for measurement without providing suitable reference points.

The relative motion of the various parts of the trail is calculated from measurements made at intervals of up to 6 or 7 minutes. The result is then the mean shear over the period.

It is advisable to arrange that the observing aircraft flies at a height rather above the bottom of the trail and on a course of a given pattern e.g. in a square around the trail. Readings are then taken at 10 seconds intervals of its air position, heading and speed and finally of the bearing of the trail relative to the aircraft's heading. This is done from a bearing plate mounted in the aircraft's fuselage near an open window. Elevation angle of any recognisable point on the trail can also be read. If this square is flown around the trail the centre of the "cats cradle" formed by the intersection of lines drawn at the horizontal bearings from the respective air positions in a plan of the square will be the location of the trail relative to the square. Hence on each photograph the distance of the trail away from the observing aircraft is known and also the azimuth at which the photograph is taken. A photograph taken on a reciprocal azimuth some time later will reveal the relative deformations that have occurred along the trail in a direction perpendicular to this azimuth. By measurement of these relative deformations over the given time interval and knowing the distance of the trail at any point from the observing aircraft the components of wind shear perpendicular to this azimuth may be calculated. If this calculation is repeated for a minimum of two well spaced azimuths the wind shears may then be found and if several azimuths are used the effect of observational errors can be reduced. In order to achieve a reasonable accuracy it is advisable to make the observations on about ten azimuths as both the aircraft's course and the bearing of the trail relative to this course can only be measured to an accuracy of about one degree. It is of course essential that each photograph should contain a horizon as a reference point for the measurements. It is not necessary to obtain ground positions on this work except to find the wind at the observing aircraft's level and then compound it with the wind shears to find the wind at the other levels.

## The experimental procedure.

The Flight Plan of the Hastings aircraft during the experiments is shown in Fig. 4. The details of operation are as follows, the two aircraft being in contact by radiotelephone and the crew of the Hastings all taking simultaneous observations at ten seconds intervals timed over the aircraft's intercommunication system:-

1. The observing aircraft flies on a steady course with the bomb dropping aircraft abeam to port, 5,000 feet higher and about 2 n.m. distant (i.e. at an elevation of 30° from the observing aircraft).

2. When the aircraft are satisfactorily positioned the bomb is dropped and as soon as the trail is fully formed timing is started and the first photograph taken.

3. Thereafter bearings on the bottom of the trail, and photographs, are taken every ten seconds. A blank is exposed between each fifty seconds and the full minute.

4. At +30 sec. a turn is made through 90° to port and completed by 1 min.00 sec. The aircraft then flies straight and level for 1 min. on the new heading and then turns through 90° to port again, completing the turn by 2 min. 30 sec. and flying straight and level for a further 1 min. This procedure is continued until a complete square has been flown around the trail.

5. Air positions, speed, height, and aircraft heading are also taken every 10 seconds around the square.

6. By taking a 'Gee' fix (ground) and an air position fix before and after the square is flown a wind is obtained for the level of the observing aircraft. This wind is the average wind over the square at the level of the aircraft.

/Analysis.

## Analysis.

1. The A.Ps. are plotted to a suitable scale. Air positions are given direct in degrees and minutes of latitude and longitude so that the necessary conversions must be applied to give a plot in nautical miles. A device called the A.P.I. wind finding attachment is available which reads to seconds. The A.Ps. are only true latitude and longitude if the correct reading is set at base and the aircraft flies in zero wind. The plot of A.Ps. should be consistent with a plot calculated from aircraft speed and headings.

2. The bearings of the bottom of the trail or preferably the portion at the observing aircraft's level, if it can be recognised, are plotted and its position is fixed at the point of best intersection. The photographs are then paired off along reciprocal bearings (Fig. 3).

3. The points down the trail at which the deformation due to wind shear is most recognisable are chosen. (Fig. 3).

4. The distance of the trail from the observing aircraft is measured for each photograph and the equivalents, in yards, of the various distances on each photograph are calculated.

5. Using a transparent grid, divided into small squares, the development both horizontally and vertically of each point relative to a point in the trail on a level with the aircraft is obtained, and is expressed in yards.

6. The development between photographs on reciprocal bearings is calculated, from which, knowing the time between them, a component of the speed is obtained in the direction of the plane of the photographs.

7. These components are plotted for each point. The best intersection of the lines drawn at right angles to each vector will give the wind relative to the wind at the level of the aircraft. An example is given in Fig. 5.

8. The relative winds are then drawn on the A.P. plot and the distance of each point from the aircraft at each ten seconds is remeasured.

9. A redetermination of the relative winds is then carried out as before. This reduces the errors resulting from a neglect of the drift of different portions of the trail in the first determinations.

10. To obtain the true winds the wind at the level of the aircraft is added vectorially to each relative wind.

## Analysis of a smoke trail dropped on 10/10/52.

Conditions were smooth throughout this experiment which was of course primarily designed to establish a technique of measurement, and in fact none of this series was dropped in turbulent conditions.

1. In this experiment the photographs were taken with a standard R.A.F. F.24 camera with a focal length of 8 in., giving a picture 5 in. x 5 in. This was enlarged four times for the purposes of analysis.

A complication in this case was that the point on the trail level with the aircraft was not recognisable from photograph to photograph for the first 50 seconds. The best fix of the position of the trail was found later at about +2 min. by taking bearings on the middle of the trail.

Seven points on the trail were chosen where the direction of shear appeared to change (Fig. 1). They were chosen from the photograph at 2 min. 40 sec. and traced back to +40 sec. and on to 5 min. 30 sec. after which they could no longer be recognised.

Point No. 1 was near the top of the trail and point No. 7 on a level with the observing aircraft.

2. The analysis proceeded similarly to the general case already described except that the winds were related to point 1 as zero because point 7 was not recognisable at + 40 sec. and + 50 sec. By adding vectorially the reciprocal of the relative wind at point 7 to the winds at the other points the winds relative to point 7 as zero were obtained.

3. To find the position of the trail at zero time the reciprocal vector representing the distance covered in 2 min. by the trail at point 3 (the point fixed by bearings at about 2 min.) relative to point 7 (i.e. relative to the air position plot) was drawn from the fixed position.

4. All the wind vectors relative to point 7 as zero were drawn from the end of the reciprocal vector, and the distance covered in each 10 seconds was marked off.

5. The distances of the points from the aircraft at each 10 seconds were remeasured and the winds redetermined (Fig. 5). The actual wind at point 7 was added vectorially to the relative winds to give the true winds at all points. The relative and true winds are given below.

| Point. | Mean Height. | Wind relative to air position plot as zero. | True Wind. |
|---|---|---|---|
| 1. | 17850 | 308°/15.5 kt. | 314°/55.5 kt. |
| 2. | 16980 | 325°/15.0 " | 318°/55.0 " |
| 3. | 16730 | 334°/15.5 " | 321°/55.0 " |
| 4. | 16380 | 338°/11.5 " | 321°/51.0 " |
| 5. | 16160 | 358°/ 8.5 " | 323°/47.0 " |
| 6. | 16030 | 012°/ 8.5 " | 325°/45.0 " |
| 7. | 15000 | 0 | 316°/40.0 " |

## Discussion.

The relative winds are estimated to be correct to within ± 5° in direction and ± ½ knot in speed.

This range of error should allow a good estimation of wind shear to be obtained and the method of marking the atmosphere with smoke allows the proper points to be used in determining the shear.

The absolute value of the true wind speed determined from the aircraft is subject to the error inherent in the navigational instruments. A wind determined by 'Gee' fixes and air positions taken 10 minutes apart is subject to an error of about ± 5° in direction and ± 5 knots in speed.

In this analysis no account has been taken of vertical motion although attempts were made to obtain measurements of the vertical movements of the different points relative to each other.

Systematic changes of height of the points apparently existed. Disregarding the measurements made after + 5 min. when identification of points on the trail from photograph to photograph may have been considerably in error it appears from the other measurements that vertical motion of as much as 300 ft./min. was present at times. This is possibly an overestimate as the reference point was not necessarily stationary in height and may be moving in the same or the opposite vertical sense.

It was not possible of course to obtain a reference point fixed in height.

A feature of this method of finding winds is that the wind is obtained at a given point in the atmosphere. It is a true spot wind and not a mean wind over a layer.

Although the analysis is tedious it is not difficult and the method can be applied to any problem that requires accurate details of the wind field in clear air.

## Conclusions and recommendations.

As a means for producing a vertical smoke trail this generator is satisfactory up to a height of about 25,000 feet. At greater altitudes the trail becomes too tenuous and disperses too quickly. It is therefore recommended that for work at higher altitudes a rather larger generator with a faster rate of burning be designed. It is very desirable to provide any further generators with built in delayed action fuses set to say 2 or 3 sec. delay. A standard bomb mounting lug (e.g. type SIS 717) should be built into the bomb and then standard carriers and electrical release mechanisms could be used in future work. To ensure a consistent ballistic performance it would be an advantage to fit a cruciform tail unit on to the bomb and also to round its nose slightly.

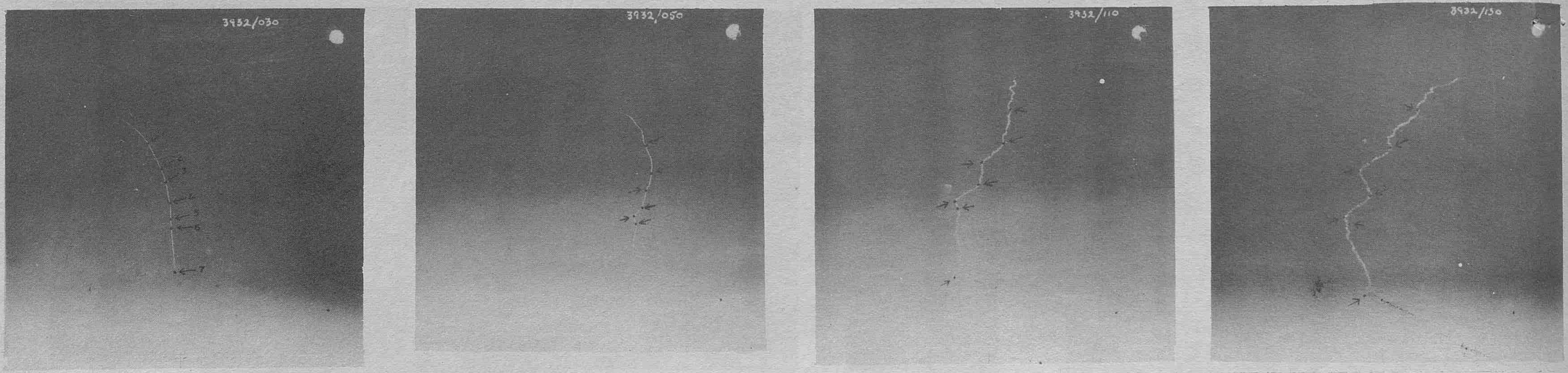
It would simplify the measurement and analysis problem considerably if the bomb could be made to emit separate puffs of smoke at intervals as it falls in addition to its continuous burning. This would provide reference points which would help considerably in assessing the likelihood of vertical motions and the consistency of the reading of crests and troughs along the trail.

As a tool in investigating clear air turbulence the method suffers from the difficulty that it is a set piece experiment requiring the presence of two specially equipped aircraft and the occasions when all conditions are suitable for the experiment will be very few. It is probably worthwhile however to conduct a limited number of experiments producing trails in the vicinity of jet streams and through the tropopause and it is recommended that about twenty more of these generators suitably modified be supplied to the Meteorological Research Flight for this purpose.

Table 1.

Summary of Experiments.

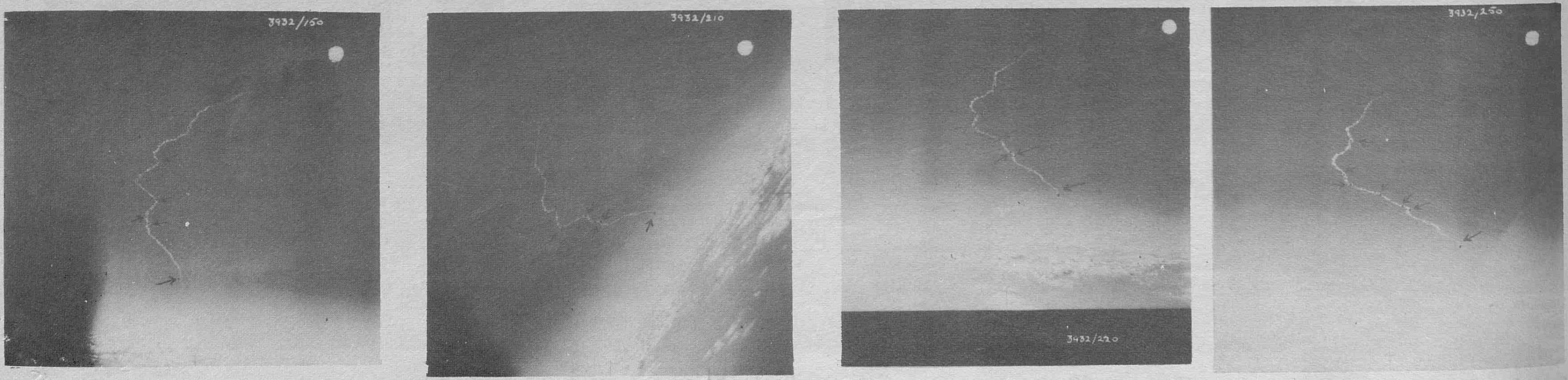
| No. | Date | Height of top (feet) | Approx. height of bottom (feet) | Height of Hastings (feet) | Effective duration (minutes) | Colour | Background | Remarks |
|---|---|---|---|---|---|---|---|---|
| 1. | 25/6/52 | 20,000 | 15,000 | 15,000 | 18 visible.<br>7 photographically | White | Cirrus | Photographs were taken but contrast was poor and no analysis was possible. Smooth ballistic curve at first quickly became a spiral shape. |
| 2. | 11/7/52 | 25,000 | 19,000 | 20,000 | 10 visible | White | Cirrus | Similar. |
| 3. | 16/7/52 | 30,000 | 23,000 | 22,000 | 7 visible | White | Blue sky | Attempts to fly given Flight Plan not successful. |
| 4. | 10/10/52 | 20,000 | 15,000 | 15,000 | 6 photographically | Intense white | Blue sky | Most successful experiment. Described in detail in report. |
| 5. | 17/10/52 A.M. | 25,000 | - | - | - | - | - | Generator did not ignite. |
| 6. | 17/10/52 P.M. | 25,000 | 17,500 | 15,000 | 5 photographically | White | Some Cirrus | Camera fault after $3\frac{1}{2}$ minutes. Partly successful experiment but not enough data for successful analysis. |

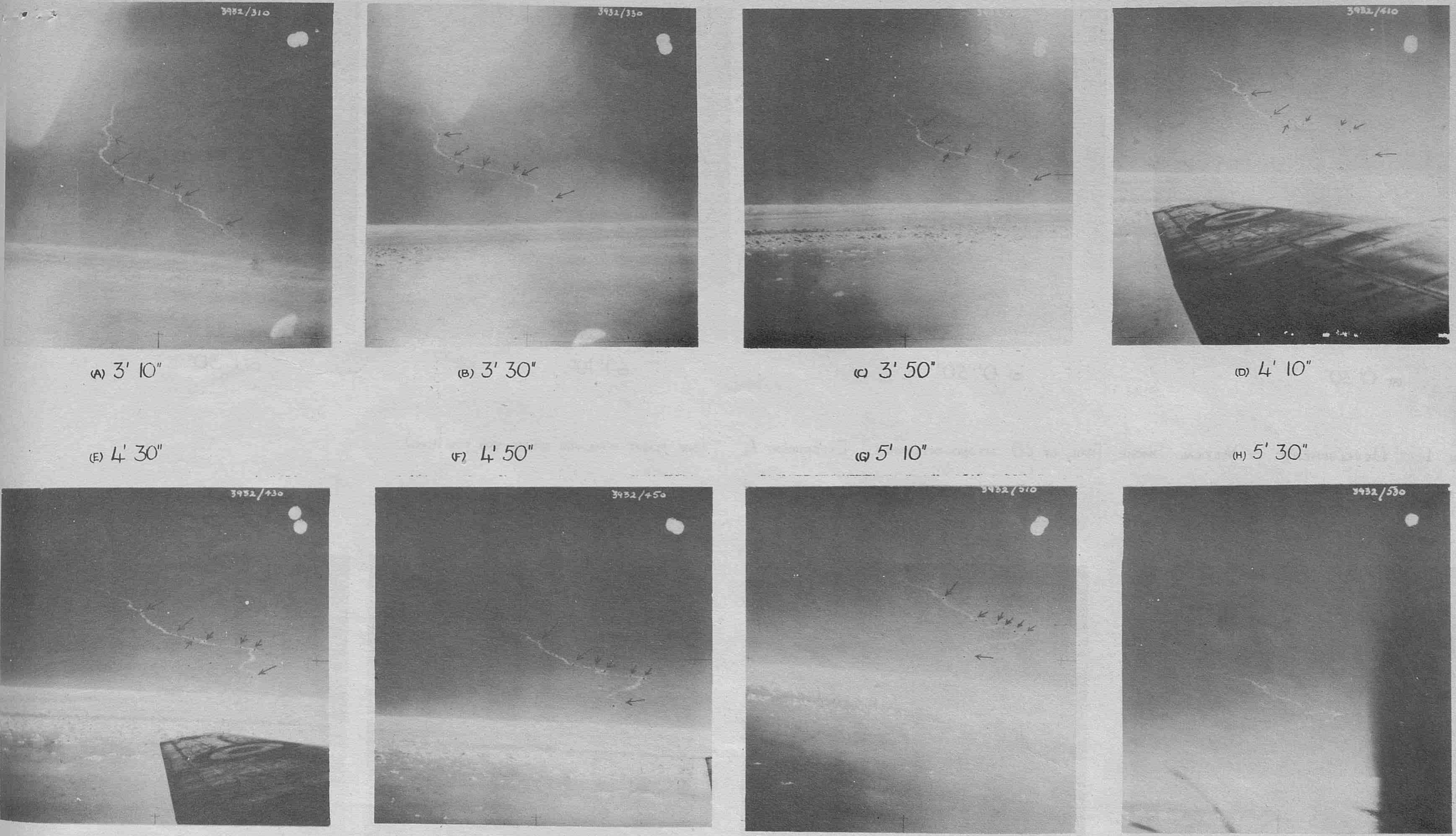

3932/030
3932/050
3932/110
3932/130

(A) 0' 30"  (B) 0' 50"  (C) 1' 10"  (D) 1' 30"

Figure 1 Development of Vertical Smoke Trail, at 20 second intervals, Experiment 4 (see flight plan for times and positions)

(E) 1' 50"  (F) 2' 10"  (G) 2' 30"  (H) 2' 50"

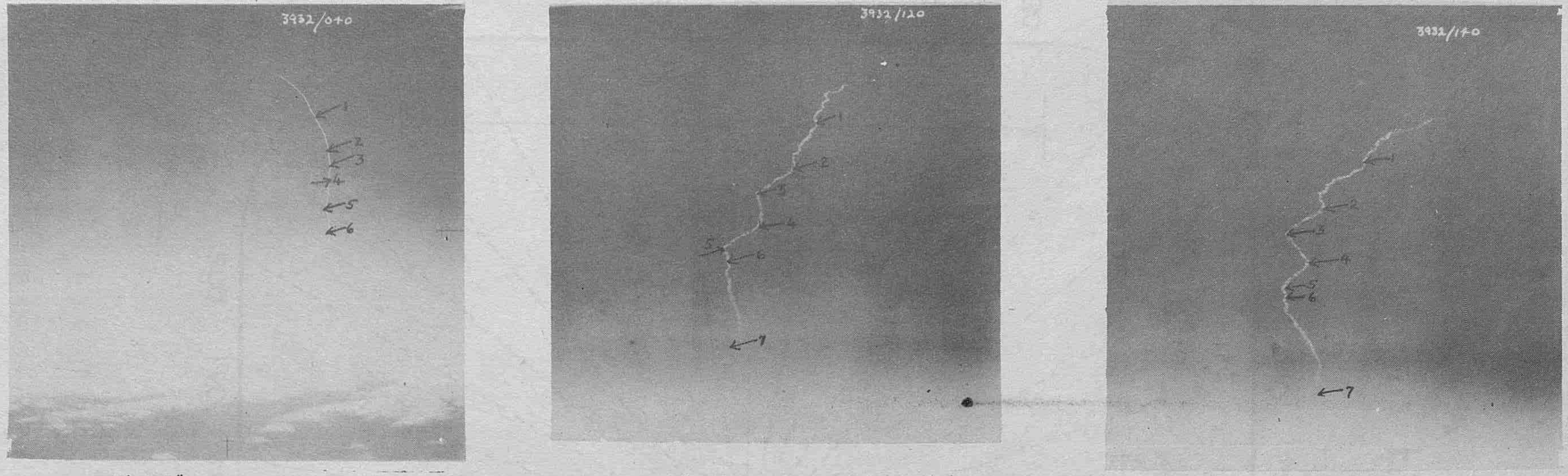

3932/150
3932/210
3432/220
3932/250



3932/310
3932/330
3932/410
3932/430
3932/450
3932/510
3932/530

(A) 3' 10"
(B) 3' 30"
(C) 3' 50"
(D) 4' 10"
(E) 4' 30"
(F) 4' 50"
(G) 5' 10"
(H) 5' 30"

FIGURE 2 DEVELOPMENT OF VERTICAL SMOKE TRAIL AT 20 SECOND INTERVALS, EXPERIMENT 4 (SEE FLIGHT PLAN FOR TIMES AND POSITIONS)


3932/040
1
2
3
4
5
6
3932/120
1
2
3
4
5
6
7
3932/140
1
2
3
4
5
6
7

A (i) 0' 40"

B (i) 1' 20"

C (i) 1' 40"

AZIMUTH RIGHT TO LEFT
350°

AZIMUTH RIGHT TO LEFT
308°

AZIMUTH RIGHT TO LEFT
270°

A (ii) 2' 40"

B (ii) 3' 40"

C (ii) 4' 40"

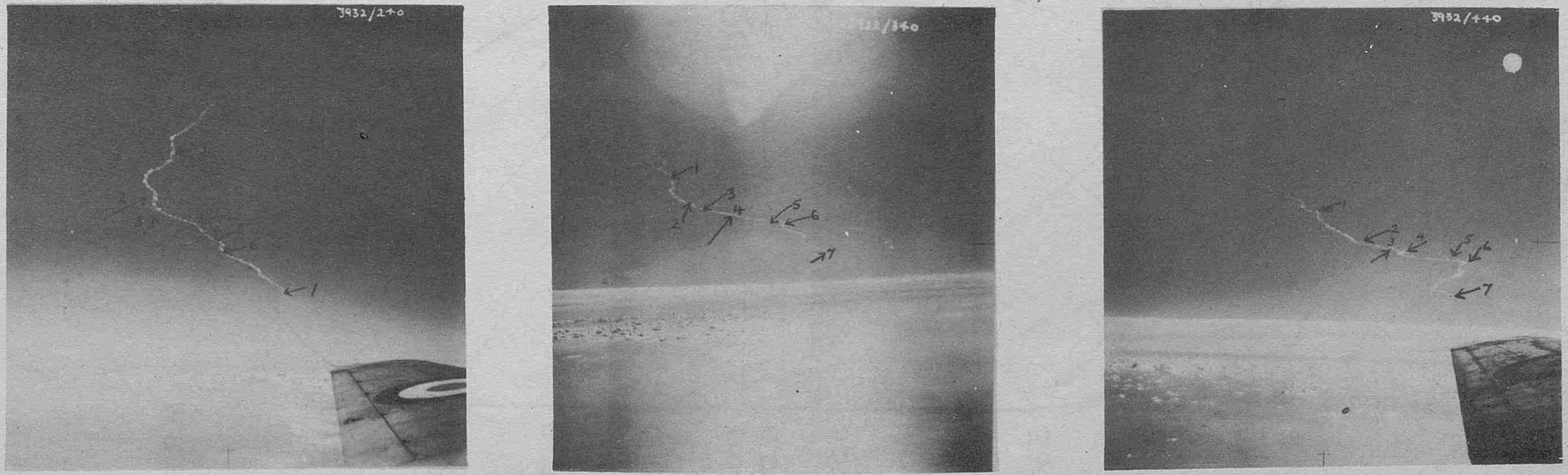

3932/240
1
2
3
4
5
6
3932/340
1
2
3
4
5
6
7
3932/440
1
2
3
4
5
6
7

FIGURE 3. THREE TYPICAL RECIPROCAL PAIRS, SHEWING POINTS CHOSEN

M.R.P. 810.

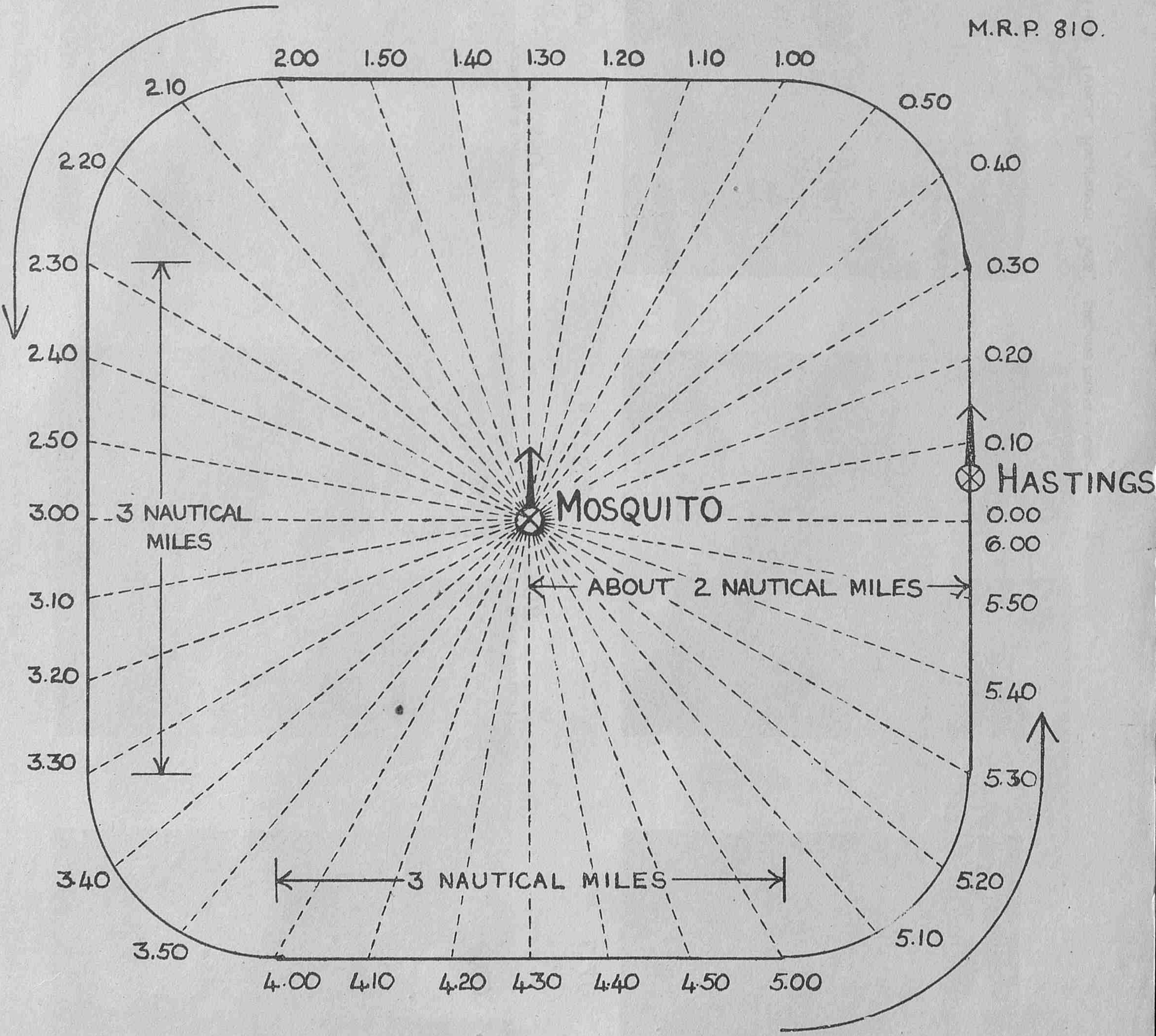
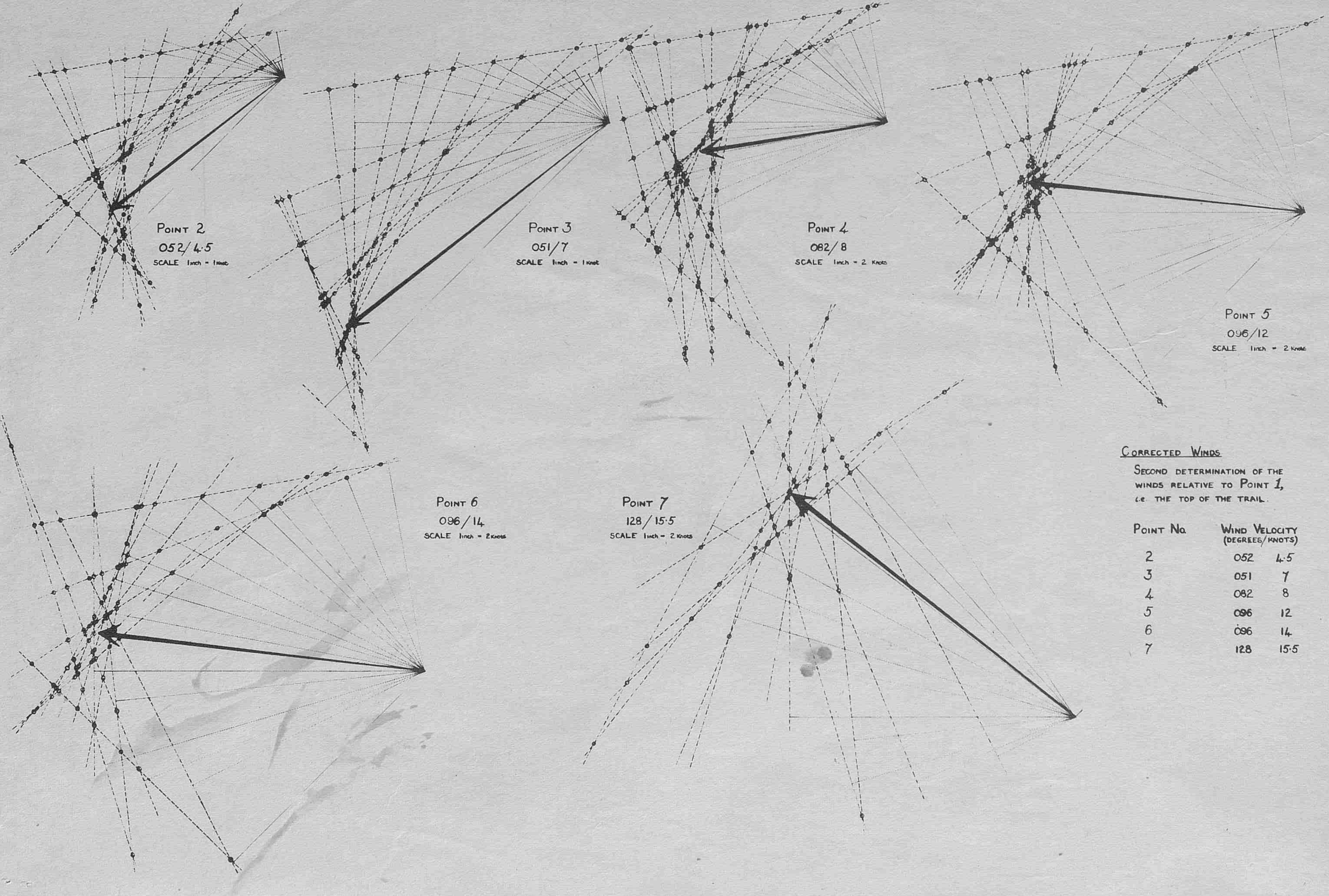

2.00
1.50
1.40
1.30
1.20
1.10
1.00
2.10
0.50
2.20
0.40
2.30
0.30
2.40
0.20
2.50
0.10
HASTINGS
3.00
3 NAUTICAL MILES
MOSQUITO
0.00
6.00
ABOUT 2 NAUTICAL MILES
3.10
5.50
3.20
5.40
3.30
5.30
3.40
3 NAUTICAL MILES
5.20
3.50
5.10
4.00
4.10
4.20
4.30
4.40
4.50
5.00

FIGURE 4

Figure 5 : Corrected Winds Relative to Point 1



Point 2
052/4·5
SCALE 1inch = 1 knot

Point 3
051/7
SCALE 1inch = 1 knot

Point 4
082/8
SCALE 1inch = 2 Knots

Point 5
096/12
SCALE 1inch = 2 Knots

Point 6
096/14
SCALE 1inch = 2 knots

Point 7
128/15·5
SCALE 1inch = 2 knots

Corrected Winds

Second determination of the winds relative to Point 1, i.e. the top of the trail.

| Point No. | Wind Velocity (degrees/knots) | |
|---|---|---|
| 2 | 052 | 4·5 |
| 3 | 051 | 7 |
| 4 | 082 | 8 |
| 5 | 096 | 12 |
| 6 | 096 | 14 |
| 7 | 128 | 15·5 |

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 811 by Miss E.E. Austin and Mr. D. Dewar.

A copy is available in the Library of the Meteorological Office."

M.R.P. 811
[illegible]
18th May, 1951.

AD 37927

AIR TE[illegible] TECHNICAL [illegible]

## AIR MINISTRY
## METEOROLOGICAL RESEARCH COMMITTEE

# Upper winds over the Mediterranean and Middle East

by

E.E. Austin and D. Dewar (Meteorological Office)

During the routine summarising of data in M.O.23, the Upper Air Climatology Branch, vector resultant winds for the period 1948-50 have been computed for all stations under the control of the Meteorological Office in the Mediterranean and Middle East and also for Nairobi, data for which are received by the courtesy of the Director of the British East African Meteorological Service. Most of these data will be published eventually station by station in M.O.555 "Upper Air Data", but in view of the importance of the Middle East in civil aviation it is thought that it may be of value to issue a summary in advance of the routine publication.

The co-ordinates of the stations are:-

| | | | |
|---|---|---|---|
| (a) | Gibraltar | 36°09'N. | 05°21'W. |
| (b) | Malta | 35°50'N. | 14°27'E. |
| (c) | Benina | 32°05'N. | 20°16'E. |
| (d) | Nicosia | 35°09'N. | 33°17'E. |
| (e) | Habbaniya | 33°22'N. | 43°34'E. |
| (f) | Bahrein | 26°16'N. | 50°37'E. |
| (g) | Aden | 12°49'N. | 45°02'E. |
| (h) | Nairobi | 01°17'S. | 36°50'E. |

Two observations a day at 0300 and 1500 G.M.T. were made at all stations except Nairobi, at which there was only one, at 0600 G.M.T.

The computed data give the mean vector resultant wind, the standard vector deviation and the mean scalar speed at the levels 850, 700, 500, 300, 200, 150 and 100 mb. in each month. Owing to the high level of the station at Nairobi data there are not available for 850 mb.

The resultant winds at any level were computed by grouping the individual observations into 30° ranges of direction, and resolving the total speed in each of these twelve 30° ranges into N-S. and E-W. components.

In computing the standard vector deviations the sum of the squares of the individual speeds was obtained by grouping the observations into 10 or 20 knot ranges of speed and multiplying the frequency in each range by the square of the mean speed of the range.

No attempt has been made to correct for the loss of observations at high levels on occasions of strong wind. This loss is likely to be serious in the winter months when at some stations speeds of over 100 knots are frequent and the proportion of balloons lost is relatively high. As a guide to the amount by which the winds at high levels are biassed, the number of observations is included in the tables.

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M.R.P. 811
S.C.II/148
18th May, 1953.

AD 37927

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The data are set out in Tables 1-8 and are represented pictorially in three different forms. Figs. 1(a) - (h) illustrate the annual variation of the vector resultant at each of the eight stations. Figs. 2(a) - (d) show the same data plotted on monthly charts to illustrate the resultant flow over the Middle East at levels from 500 mb. upwards; Figs. 3(a) - (d) show the variation with height of the mean vector wind speed.

/Among

G.366514/MKL/5/53/70

Among many interesting features shown in Figs. 1 and 2 the following may be noted.

(i) The strong westerly winds in winter - the subtropical jet-stream. The axis of the jet is in about the latitude of Bahrein (Fig. 2(a)), and the winds attain their greatest speed in January and February between 200 and 150 mb.

(ii) The marked falling off in the speed of the wind at Bahrein (Fig.1(f)) between April and May. At 200 mb. the highest speeds in May occur at Benina and Habbaniya, and in June at Nicosia (Fig. 2). It is noteworthy that at Nicosia the speed in June is as great as in January (Fig. 1(d)).

(iii) The temporary strengthening of the wind at Nicosia (Fig. 1(d)) in September and at Habbaniya (Fig. 1(e)) in October.

(ii) and (iii) indicate that at 200 mb. the zone of strongest wind moves northwards in summer and retreats southwards again in autumn. A strengthening of the westerlies at 200 mb. over Malta in April (Fig. 1(b)) and their persistence throughout the summer may be another aspect of the same phenomenon. At Gibraltar a similar increase in speed is shown in May but is followed there by a gradual decrease.

(iv) The strong easterly winds at Aden (Fig. 1(g)) from 200 mb. upwards from June to September (M.R.P. 729) and the occurrence of easterlies at Bahrein (Fig. 1(f)) at the same levels but lighter in speed and of shorter duration.

(v) The well-marked southerly component of the westerly winds at high levels in summer at Benina, Nicosia and Habbaniya (Figs. 2(b) and (c)).

(vi) The intensification of the easterly wind at Nairobi (Fig. 1(h)) in January and February particularly at 200 mb. and in July at 150 mb.

Fig. 3 is, of course, affected by the loss of balloons at high levels in strong winds, and the marked falling off in speed above 200 mb. in winter is likely to be over-emphasised. Figs. 3(a) and (b) are for stations from east to west across the Mediterranean between latitudes 32° and 36°N. and Figs. 3(c) and 3(d) are for stations from north to south (33°N. to 1°S.) between longitudes 37° and 51°E.

The complete change in the wind structure at both Bahrein and Aden between May and June is very clearly shown. At Bahrein the wind in the summer months continues to increase with height up to the highest level shown, and at Aden also in June and October and possibly from July to September when the losses in strong winds are allowed for.

It is hoped to continue the study by examining the day to day changes at high levels, and also the variability of the vector resultants from year to year.

## Table 1. DIRECTION AND SPEED OF UPPER WINDS

## GIBRALTAR 36°09'N. 05°21'W. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | Mean Vector Wind (knots) | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 296 | 292 | 282 | 271 | 267 | 260 | 248 | 257 | 264 | 280 | 291 | 282 |
| kt. | 18 | 21 | 23 | 21 | 23 | 17 | 14 | 14 | 10 | 12 | 19 | 21 |
| 150 ° | 301 | 292 | 279 | 269 | 264 | 262 | 256 | 261 | 261 | 295 | 293 | 285 |
| kt. | 20 | 26 | 28 | 25 | 35 | 28 | 29 | 28 | 15 | 17 | 27 | 28 |
| 200 ° | 311 | 306 | 282 | 273 | 263 | 259 | 257 | 264 | 250 | 299 | 295 | 290 |
| kt. | 19 | 21 | 26 | 21 | 41 | 30 | 31 | 27 | 17 | 17 | 29 | 32 |
| 300 ° | 321 | 327 | 286 | 278 | 257 | 251 | 253 | 259 | 236 | 286 | 294 | 292 |
| kt. | 15 | 17 | 18 | 11 | 33 | 27 | 22 | 19 | 17 | 12 | 24 | 33 |
| 500 ° | 314 | 314 | 286 | 275 | 254 | 246 | 237 | 252 | 235 | 278 | 291 | 291 |
| kt. | 13 | 10 | 9 | 7 | 23 | 21 | 20 | 18 | 15 | 9 | 17 | 21 |
| 700 ° | 300 | 297 | 270 | 260 | 246 | 234 | 235 | 239 | 228 | 269 | 285 | 280 |
| kt. | 8 | 6 | 5 | 4 | 13 | 10 | 15 | 15 | 11 | 5 | 11 | 14 |
| 850 ° | 279 | Calm | 110 | 84 | 250 | 132 | 163 | 183 | 171 | 109 | 274 | 277 |
| kt. | 5 | | 5 | 2 | 5 | 5 | 4 | 3 | 4 | 3 | 4 | 7 |
| | | | | | Standard Vector Deviation (knots) | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 18 | 21 | 19 | 19 | 18 | 18 | 21 | 16 | 16 | 16 | 21 | 21 |
| 150 | 26 | 26 | 26 | 24 | 27 | 28 | 31 | 24 | 31 | 27 | 31 | 28 |
| 200 | 34 | 37 | 34 | 32 | 35 | 35 | 35 | 28 | 39 | 36 | 38 | 37 |
| 300 | 43 | 43 | 38 | 41 | 38 | 34 | 31 | 26 | 35 | 35 | 34 | 43 |
| 500 | 33 | 29 | 27 | 30 | 27 | 24 | 23 | 17 | 23 | 23 | 25 | 30 |
| 700 | 24 | 22 | 20 | 21 | 19 | 16 | 19 | 14 | 18 | 17 | 19 | 22 |
| 850 | 21 | 22 | 20 | 19 | 18 | 17 | 16 | 14 | 19 | 16 | 18 | 22 |
| | | | | | Mean Scalar Speed (knots) | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 23 | 25 | 27 | 25 | 26 | 22 | 21 | 18 | 16 | 18 | 25 | 26 |
| 150 | 28 | 32 | 34 | 30 | 39 | 35 | 37 | 32 | 30 | 27 | 35 | 37 |
| 200 | 33 | 35 | 37 | 33 | 48 | 41 | 42 | 34 | 38 | 34 | 41 | 44 |
| 300 | 39 | 39 | 36 | 36 | 45 | 37 | 33 | 27 | 34 | 32 | 36 | 47 |
| 500 | 30 | 26 | 23 | 25 | 30 | 27 | 26 | 21 | 23 | 21 | 25 | 31 |
| 700 | 21 | 19 | 17 | 18 | 20 | 17 | 21 | 18 | 18 | 15 | 19 | 22 |
| 850 | 18 | 19 | 18 | 17 | 17 | 15 | 14 | 13 | 14 | 14 | 16 | 19 |
| | | | | | Number of Observations | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 128 | 146 | 185 | 174 | 153 | 190 | 183 | 174 | 154 | 190 | 190 | 126 |
| 150 | 160 | 180 | 229 | 200 | 203 | 224 | 215 | 214 | 193 | 233 | 233 | 163 |
| 200 | 183 | 210 | 247 | 222 | 229 | 238 | 228 | 228 | 215 | 254 | 249 | 191 |
| 300 | 207 | 230 | 266 | 253 | 255 | 254 | 237 | 248 | 232 | 267 | 261 | 227 |
| 500 | 239 | 247 | 276 | 261 | 272 | 268 | 239 | 261 | 247 | 279 | 265 | 241 |
| 700 | 239 | 248 | 277 | 262 | 272 | 268 | 239 | 261 | 247 | 279 | 265 | 247 |
| 850 | 240 | 248 | 277 | 262 | 273 | 268 | 239 | 261 | 247 | 279 | 265 | 254 |

/Table 2.

## Table 2. DIRECTION AND SPEED OF UPPER WINDS

### MALTA 33°50'N. 14°27'E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mean Vector Wind (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 275 | 285 | 285 | 265 | 270 | 265 | 245 | 250 | 265 | 275 | 280 | 270 |
| kt. | 31 | 28 | 25 | 26 | 22 | 20 | 21 | 19 | 17 | 18 | 25 | 30 |
| 150 ° | 270 | 280 | 285 | 275 | 280 | 275 | 255 | 260 | 275 | 275 | 285 | 275 |
| kt. | 40 | 36 | 34 | 38 | 36 | 36 | 38 | 36 | 28 | 22 | 33 | 38 |
| 200 ° | 270 | 285 | 290 | 275 | 280 | 280 | 265 | 265 | 290 | 275 | 290 | 275 |
| kt. | 40 | 36 | 31 | 41 | 40 | 38 | 40 | 36 | 29 | 21 | 32 | 40 |
| 300 ° | 280 | 295 | 295 | 275 | 275 | 290 | 285 | 285 | 295 | 280 | 285 | 275 |
| kt. | 28 | 25 | 23 | 37 | 37 | 33 | 29 | 23 | 24 | 15 | 27 | 36 |
| 500 ° | 290 | 300 | 300 | 280 | 275 | 295 | 305 | 305 | 300 | 280 | 285 | 280 |
| kt. | 17 | 19 | 17 | 22 | 22 | 26 | 19 | 15 | 16 | 12 | 19 | 21 |
| 700 ° | 285 | 305 | 310 | 275 | 280 | 300 | 320 | 320 | 305 | 275 | 285 | 265 |
| kt. | 12 | 12 | 12 | 13 | 13 | 18 | 16 | 11 | 12 | 10 | 14 | 14 |
| 850 ° | 275 | 315 | 315 | 280 | 280 | 305 | 320 | 315 | 295 | 265 | 285 | 260 |
| kt. | 10 | 7 | 8 | 6 | 4 | 8 | 7 | 6 | 6 | 6 | 10 | 10 |
| Standard Vector Deviation (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 23 | 22 | 18 | 18 | 16 | 17 | 19 | 15 | 16 | 14 | 18 | 26 |
| 150 | 31 | 28 | 25 | 27 | 21 | 20 | 23 | 23 | 24 | 24 | 25 | 33 |
| 200 | 41 | 39 | 32 | 33 | 30 | 26 | 28 | 27 | 29 | 32 | 38 | 39 |
| 300 | 52 | 49 | 40 | 38 | 35 | 27 | 27 | 26 | 26 | 30 | 39 | 44 |
| 500 | 37 | 33 | 27 | 26 | 23 | 21 | 19 | 18 | 18 | 21 | 28 | 31 |
| 700 | 26 | 24 | 19 | 20 | 19 | 17 | 19 | 16 | 16 | 17 | 21 | 21 |
| 850 | 22 | 22 | 17 | 18 | 15 | 14 | 13 | 12 | 13 | 14 | 17 | 18 |
| Mean Scalar Speed (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 34 | 31 | 28 | 30 | 24 | 23 | 25 | 22 | 20 | 21 | 28 | 34 |
| 150 | 45 | 40 | 37 | 42 | 38 | 38 | 41 | 39 | 33 | 28 | 39 | 45 |
| 200 | 51 | 46 | 39 | 49 | 45 | 42 | 45 | 40 | 36 | 32 | 46 | 50 |
| 300 | 52 | 48 | 40 | 47 | 44 | 39 | 34 | 29 | 31 | 28 | 42 | 50 |
| 500 | 35 | 34 | 28 | 30 | 28 | 30 | 24 | 20 | 21 | 21 | 30 | 32 |
| 700 | 25 | 23 | 20 | 20 | 20 | 23 | 20 | 17 | 17 | 17 | 22 | 22 |
| 850 | 21 | 20 | 16 | 16 | 13 | 14 | 12 | 11 | 11 | 13 | 18 | 18 |
| Number of Observations | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 81 | 78 | 88 | 73 | 71 | 74 | 94 | 99 | 107 | 134 | 104 | 95 |
| 150 | 119 | 111 | 123 | 115 | 101 | 99 | 111 | 114 | 128 | 161 | 147 | 136 |
| 200 | 145 | 146 | 152 | 126 | 127 | 124 | 128 | 125 | 134 | 168 | 165 | 153 |
| 300 | 174 | 164 | 175 | 146 | 152 | 142 | 138 | 138 | 138 | 172 | 171 | 176 |
| 500 | 186 | 169 | 183 | 150 | 155 | 150 | 144 | 147 | 146 | 176 | 175 | 184 |
| 700 | 186 | 169 | 184 | 150 | 155 | 150 | 146 | 150 | 147 | 181 | 176 | 185 |
| 850 | 186 | 170 | 184 | 150 | 155 | 150 | 148 | 150 | 147 | 181 | 177 | 185 |

/Table 3.

## Table 3. DIRECTION AND SPEED OF UPPER WINDS

## BENINA 32°05'N. 20°16'E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June. | July | Aug. | Sep. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Mean Vector Wind (knots)** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 268 | 270 | 276 | 276 | 264 | 242 | 201 | 213 | 245 | 256 | 268 | 275 |
| kt. | 57 | 46 | 42 | 37 | 23 | 23 | 17 | 17 | 18 | 25 | 34 | 34 |
| 150 ° | 264 | 266 | 274 | 270 | 274 | 248 | 233 | 234 | 250 | 264 | 270 | 271 |
| kt. | 76 | 66 | 64 | 54 | 45 | 43 | 28 | 30 | 35 | 36 | 42 | 52 |
| 200 ° | 260 | 266 | 272 | 268 | 280 | 256 | 247 | 246 | 259 | 265 | 273 | 271 |
| kt. | 96 | 71 | 67 | 57 | 53 | 46 | 31 | 35 | 34 | 38 | 42 | 59 |
| 300 ° | 261 | 270 | 273 | 277 | 283 | 276 | 262 | 255 | 278 | 272 | 269 | 269 |
| kt. | 85 | 50 | 55 | 45 | 44 | 34 | 27 | 31 | 24 | 30 | 29 | 51 |
| 500 ° | 267 | 269 | 277 | 278 | 284 | 301 | 315 | 295 | 297 | 275 | 268 | 267 |
| kt. | 41 | 23 | 28 | 27 | 26 | 24 | 12 | 11 | 16 | 18 | 20 | 30 |
| 700 ° | 270 | 277 | 283 | 282 | 283 | 314 | 339 | 329 | 322 | 283 | 268 | 267 |
| kt. | 21 | 13 | 16 | 14 | 15 | 17 | 12 | 11 | 12 | 13 | 14 | 17 |
| 850 ° | 266 | 265 | 289 | 286 | 301 | 330 | 354 | 357 | 338 | 286 | 259 | 257 |
| kt. | 10 | 7 | 9 | 6 | 5 | 9 | 13 | 11 | 9 | 6 | 9 | 11 |
| **Standard Vector Deviation (knots)** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 23 | 22 | 23 | 22 | 19 | 17 | 19 | 15 | 12 | 16 | 23 | 23 |
| 150 | 22 | 29 | 25 | 26 | 28 | 25 | 23 | 22 | 23 | 24 | 25 | 35 |
| 200 | 42 | 35 | 32 | 40 | 35 | 28 | 24 | 25 | 29 | 32 | 31 | 43 |
| 300 | 45 | 40 | 41 | 38 | 34 | 26 | 23 | 22 | 28 | 29 | 33 | 44 |
| 500 | 32 | 24 | 29 | 27 | 23 | 21 | 13 | 14 | 15 | 21 | 22 | 30 |
| 700 | 22 | 19 | 20 | 19 | 15 | 18 | 13 | 13 | 14 | 18 | 18 | 21 |
| 850 | 17 | 15 | 18 | 17 | 15 | 13 | 11 | 12 | 11 | 14 | 16 | 16 |
| Surface | 10 | 11 | 12 | 12 | 13 | 11 | 10 | 12 | 8 | 11 | 11 | 10 |
| **Scalar Speed (knots)** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 58 | 47 | 44 | 40 | 26 | 25 | 21 | 19½ | 20 | 27 | 36 | 37 |
| 150 | 79½ | 69 | 66 | 58 | 49 | 46 | 33 | 34 | 38 | 39 | 46 | 57 |
| 200 | 100 | 75 | 70½ | 64 | 58 | 50 | 34 | 39 | 40 | 44 | 47 | 66 |
| 300 | 91 | 58 | 61 | 54 | 50 | 39 | 31 | 35 | 33 | 36½ | 39 | 60 |
| 500 | 46½ | 30 | 35 | 34 | 31 | 28 | 16 | 15 | 20 | 24 | 26 | 37 |
| 700 | 26 | 20 | 22 | 21 | 21 | 22 | 15½ | 14 | 16 | 19 | 20 | 23½ |
| 850 | 17 | 14½ | 17 | 16 | 13 | 14 | 16 | 15 | 13 | 13 | 16 | 18 |
| **Number of Observations** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 14 | 45 | 60 | 67 | 59 | 87 | 86 | 94 | 91 | 95 | 111 | 64 |
| 150 | 19 | 58 | 86 | 87 | 80 | 106 | 99 | 100 | 102 | 105 | 130 | 89 |
| 200 | 35 | 68 | 103 | 98 | 94 | 110 | 104 | 109 | 108 | 118 | 131 | 105 |
| 300 | 47 | 72 | 124 | 105 | 103 | 114 | 113 | 113 | 115 | 123 | 133 | 114 |
| 500 | 48 | 75 | 130 | 106 | 107 | 122 | 116 | 122 | 124 | 132 | 137 | 116 |
| 700 | 48 | 77 | 135 | 109 | 111 | 125 | 118 | 122 | 125 | 134 | 137 | 116 |
| 850 | 48 | 79 | 141 | 113 | 112 | 125 | 119 | 122 | 125 | 138 | 137 | 116 |

/Table 4.

## Table 4. DIRECTION AND SPEED OF UPPER WINDS

## NICOSIA 35°09'N. 33°17'E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | Mean Vector Wind (knots) | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 270 | 270 | 265 | 270 | 240 | 240 | 220 | 230 | 245 | 250 | 265 | 270 |
| kt. | 45 | 44 | 41 | 27 | 24 | 38 | 23 | 23 | 34 | 34 | 32 | 36 |
| 150 ° | 275 | 265 | 265 | 270 | 250 | 245 | 235 | 240 | 245 | 250 | 260 | 270 |
| kt. | 62 | 60 | 55 | 35 | 36 | 61 | 38 | 37 | 58 | 50 | 39 | 48 |
| 200 ° | 275 | 270 | 265 | 270 | 255 | 245 | 240 | 245 | 245 | 250 | 260 | 270 |
| kt. | 70 | 58 | 57 | 38 | 41 | 71 | 42 | 40 | 65 | 48 | 39 | 53 |
| 300 ° | 275 | 275 | 265 | 270 | 260 | 250 | 245 | 250 | 245 | 255 | 260 | 270 |
| kt. | 59 | 43 | 44 | 33 | 30 | 52 | 37 | 37 | 50 | 35 | 31 | 42 |
| 500 ° | 275 | 280 | 260 | 265 | 265 | 260 | 260 | 260 | 255 | 260 | 250 | 265 |
| kt. | 31 | 28 | 25 | 19 | 19 | 24 | 14 | 15 | 21 | 19 | 18 | 22 |
| 700 ° | 270 | 280 | 260 | 250 | 260 | 270 | 300 | 280 | 260 | 265 | 240 | 255 |
| kt. | 17 | 16 | 14 | 8 | 9 | 12 | 5 | 5 | 7 | 9 | 10 | 11 |
| 850 ° | 265 | 280 | 255 | 235 | 265 | 295 | 330 | 310 | 310 | 270 | 230 | 235 |
| kt. | 6 | 8 | 5 | 2 | 2 | 4 | 4 | 3 | 2 | 2 | 2 | 3 |
| | | | | | Standard Vector Deviation (knots) | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 22 | 24 | 20 | 18 | 17 | 24 | 19 | 19 | 21 | 21 | 18 | 24 |
| 150 | 32 | 32 | 24 | 22 | 25 | 29 | 24 | 24 | 25 | 27 | 26 | 34 |
| 200 | 44 | 40 | 28 | 29 | 32 | 30 | 26 | 21 | 30 | 33 | 32 | 43 |
| 300 | 51 | 45 | 35 | 37 | 34 | 31 | 27 | 23 | 29 | 34 | 34 | 43 |
| 500 | 32 | 34 | 25 | 25 | 20 | 21 | 18 | 18 | 19 | 21 | 23 | 27 |
| 700 | 20 | 21 | 19 | 17 | 17 | 16 | 12 | 12 | 13 | 15 | 17 | 19 |
| 850 | 15 | 15 | 16 | 14 | 11 | 11 | 10 | 10 | 9 | 10 | 11 | 14 |
| | | | | | Mean Scalar Speed (knots) | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 47 | 46 | 42 | 30 | 26 | 40 | 27 | 27 | 36 | 37 | 34 | 39 |
| 150 | 66 | 63 | 57 | 38 | 40 | 64 | 40 | 40 | 60 | 53 | 43 | 52 |
| 200 | 76 | 65 | 60 | 43 | 47 | 75 | 44 | 43 | 67 | 53 | 45 | 59 |
| 300 | 70 | 56 | 50 | 42 | 38 | 58 | 41 | 40 | 54 | 43 | 39 | 53 |
| 500 | 38 | 38 | 31 | 26 | 24 | 28 | 20 | 20 | 25 | 25 | 24 | 30 |
| 700 | 23 | 23 | 21 | 17 | 16 | 17 | 11 | 11 | 13 | 15 | 16 | 19 |
| 850 | 14 | 15 | 14 | 12 | 9 | 10 | 9 | 8 | 8 | 8 | 9 | 12 |
| | | | | | Number of Observations | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 26 | 47 | 60 | 74 | 86 | 79 | 101 | 101 | 89 | 138 | 130 | 106 |
| 150 | 58 | 70 | 87 | 105 | 120 | 127 | 128 | 116 | 122 | 156 | 146 | 142 |
| 200 | 97 | 103 | 114 | 131 | 135 | 145 | 142 | 123 | 137 | 161 | 157 | 155 |
| 300 | 118 | 116 | 136 | 146 | 147 | 160 | 156 | 138 | 148 | 168 | 163 | 168 |
| 500 | 133 | 128 | 145 | 151 | 154 | 165 | 169 | 158 | 153 | 171 | 163 | 170 |
| 700 | 134 | 129 | 144 | 153 | 155 | 165 | 169 | 158 | 153 | 171 | 163 | 170 |
| 850 | 134 | 129 | 143 | 153 | 155 | 165 | 169 | 158 | 153 | 171 | 163 | 170 |

/Table 5.

## Table 5. DIRECTION AND SPEED OF UPPER WINDS

## HABBANIYA $33^{\circ}22'$N. $43^{\circ}34'$E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | | Mean Vector Wind (knots) | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 275 | 265 | 270 | 265 | 240 | 240 | 190 | 215 | 245 | 260 | 270 | 270 |
| kt. | 46 | 44 | 47 | 34 | 28 | 24 | 10 | 10 | 26 | 35 | 36 | 41 |
| 150 ° | 275 | 270 | 270 | 265 | 245 | 250 | 225 | 245 | 250 | 255 | 265 | 270 |
| kt. | 65 | 67 | 72 | 51 | 48 | 46 | 14 | 18 | 42 | 59 | 48 | 56 |
| 200 ° | 275 | 270 | 265 | 260 | 250 | 250 | 240 | 255 | 255 | 255 | 265 | 270 |
| kt. | 73 | 77 | 85 | 58 | 54 | 54 | 14 | 21 | 47 | 64 | 50 | 68 |
| 300 ° | 275 | 265 | 260 | 260 | 250 | 255 | 245 | 260 | 255 | 255 | 265 | 270 |
| kt. | 63 | 68 | 70 | 48 | 44 | 48 | 14 | 20 | 42 | 48 | 39 | 59 |
| 500 ° | 270 | 265 | 260 | 255 | 250 | 265 | 230 | 265 | 260 | 260 | 260 | 265 |
| kt. | 33 | 35 | 35 | 28 | 27 | 23 | 8 | 10 | 22 | 25 | 18 | 31 |
| 700 ° | 265 | 260 | 255 | 260 | 260 | 285 | 280 | 285 | 280 | 265 | 255 | 265 |
| kt. | 14 | 17 | 18 | 14 | 16 | 15 | 7 | 9 | 15 | 18 | 10 | 14 |
| 850 ° | 275 | 260 | 260 | 255 | 295 | 315 | 310 | 310 | 320 | 280 | 250 | 260 |
| kt. | 7 | 8 | 7 | 4 | 6 | 14 | 14 | 12 | 10 | 7 | 3 | 4 |
| | | | | | | Standard Vector Deviation (knots) | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 26 | 21 | 20 | 23 | 18 | 19 | 15 | 16 | 20 | 19 | 22 | 22 |
| 150 | 31 | 29 | 28 | 29 | 24 | 25 | 19 | 20 | 26 | 28 | 26 | 34 |
| 200 | 41 | 37 | 37 | 37 | 31 | 28 | 19 | 21 | 24 | 30 | 36 | 45 |
| 300 | 48 | 44 | 37 | 38 | 33 | 28 | 19 | 20 | 23 | 27 | 36 | 43 |
| 500 | 31 | 30 | 23 | 25 | 21 | 17 | 15 | 15 | 17 | 18 | 20 | 30 |
| 700 | 19 | 20 | 17 | 19 | 19 | 15 | 15 | 15 | 14 | 14 | 14 | 17 |
| 850 | 15 | 18 | 16 | 16 | 15 | 13 | 11 | 11 | 11 | 13 | 12 | 13 |
| | | | | | | Mean Scalar Speed (knots) | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 49 | 46 | 49 | 37 | 29 | 27 | 16 | 17 | 29 | 37 | 39 | 44 |
| 150 | 69 | 69 | 74 | 55 | 51 | 48 | 19 | 23 | 45 | 61 | 51 | 61 |
| 200 | 78 | 81 | 89 | 62 | 58 | 56 | 20 | 25 | 50 | 67 | 54 | 75 |
| 300 | 71 | 75 | 74 | 55 | 49 | 51 | 20 | 23 | 44 | 50 | 45 | 67 |
| 500 | 40 | 41 | 38 | 33 | 31 | 26 | 15 | 16 | 24 | 27 | 22 | 37 |
| 700 | 20 | 23 | 22 | 21 | 22 | 19 | 15 | 16 | 19 | 21 | 15 | 19 |
| 850 | 14 | 17 | 15 | 14 | 14 | 17 | 17 | 15 | 13 | 13 | 11 | 12 |
| | | | | | | Number of Observations | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 60 | 52 | 58 | 97 | 109 | 112 | 115 | 118 | 128 | 112 | 120 | 88 |
| 150 | 109 | 93 | 120 | 133 | 133 | 134 | 135 | 134 | 153 | 149 | 136 | 119 |
| 200 | 146 | 120 | 156 | 157 | 150 | 149 | 144 | 138 | 163 | 159 | 150 | 154 |
| 300 | 181 | 144 | 173 | 174 | 172 | 169 | 151 | 141 | 169 | 164 | 160 | 178 |
| 500 | 187 | 147 | 173 | 175 | 177 | 177 | 155 | 147 | 174 | 167 | 162 | 183 |
| 700 | 187 | 147 | 173 | 176 | 177 | 177 | 159 | 147 | 175 | 167 | 162 | 184 |
| 850 | 187 | 147 | 173 | 176 | 177 | 177 | 160 | 147 | 175 | 167 | 163 | 184 |

/Table 6.

## Table 6. DIRECTION AND SPEED OF UPPER WINDS

## BAHREIN 26°16'N. 50°37'E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Mean Vector Wind (knots)** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 274 | 259 | 276 | 262 | 228 | 107 | 95 | 100 | 121 | 250 | 266 | 272 |
| kt. | 58 | 56 | 48 | 40 | 17 | 14 | 37 | 27 | 12 | 15 | 41 | 50 |
| 150 ° | 279 | 270 | 276 | 258 | 247 | 141 | 99 | 107 | 154 | 254 | 266 | 270 |
| kt. | 92 | 89 | 72 | 67 | 38 | 4 | 24 | 17 | 6 | 32 | 63 | 74 |
| 200 ° | 278 | 270 | 273 | 260 | 248 | 240 | 98 | 108 | 221 | 253 | 267 | 271 |
| kt. | 102 | 103 | 88 | 75 | 47 | 2 | 17 | 10 | 3 | 37 | 63 | 80 |
| 300 ° | 275 | 267 | 268 | 259 | 254 | 329 | 84 | 88 | 322 | 262 | 268 | 266 |
| kt. | 86 | 96 | 81 | 61 | 30 | 5 | 12 | 4 | 2 | 30 | 51 | 67 |
| 500 ° | 271 | 270 | 264 | 261 | 258 | 354 | 98 | 56 | 42 | 280 | 272 | 268 |
| kt. | 49 | 55 | 48 | 31 | 17 | 7 | 6 | 6 | 5 | 12 | 22 | 37 |
| 700 ° | 271 | 270 | 264 | 262 | 283 | 317 | 322 | 341 | 329 | 352 | 309 | 271 |
| kt. | 22 | 26 | 24 | 19 | 13 | 15 | 9 | 5 | 5 | 6 | 5 | 17 |
| 850 ° | 281 | 289 | 275 | 275 | 314 | 332 | 331 | 337 | 341 | 345 | 342 | 289 |
| kt. | 9 | 12 | 10 | 9 | 11 | 20 | 20 | 14 | 13 | 10 | 4 | 7 |
| **Standard Vector Deviation (knots)** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 23 | 34 | 16 | 23 | 16 | 20 | 19 | 18 | 15 | 19 | 25 | 31 |
| 150 | 27 | 29 | 36 | 34 | 30 | 26 | 16 | 15 | 16 | 23 | 27 | 37 |
| 200 | 34 | 39 | 45 | 38 | 38 | 26 | 13 | 13 | 18 | 28 | 34 | 43 |
| 300 | 37 | 35 | 40 | 33 | 27 | 19 | 12 | 12 | 14 | 23 | 31 | 39 |
| 500 | 29 | 30 | 27 | 23 | 19 | 13 | 16 | 12 | 15 | 18 | 18 | 25 |
| 700 | 18 | 21 | 18 | 17 | 19 | 17 | 17 | 13 | 17 | 14 | 14 | 16 |
| 850 | 14 | 16 | 16 | 15 | 19 | 15 | 14 | 14 | 13 | 11 | 12 | 14 |
| **Mean Scalar Speed (knots)** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 59 | 58 | 51 | 42 | 21 | 21 | 38 | 30 | 17 | 20 | 43 | 54 |
| 150 | 94 | 91 | 77 | 71 | 41 | 21 | 26 | 20 | 15 | 36 | 65 | 80 |
| 200 | 106 | 107 | 94 | 81 | 53 | 19 | 19 | 14 | 15 | 40 | 68 | 87 |
| 300 | 91 | 101 | 84 | 64 | 36 | 16 | 15 | 10 | 12 | 34 | 55 | 73 |
| 500 | 52 | 58 | 51 | 34 | 21 | 14 | 14 | 12 | 14 | 18 | 25 | 40 |
| 700 | 25 | 29 | 27 | 22 | 20 | 19 | 15 | 12 | 15 | 13 | 13 | 20 |
| 850 | 15 | 18 | 16 | 15 | 19 | 22 | 22 | 17 | 16 | 13 | 11 | 14 |
| **Number of Observations** | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 24 | 3 | 29 | 51 | 75 | 97 | 107 | 120 | 131 | 144 | 112 | 74 |
| 150 | 61 | 28 | 52 | 80 | 90 | 105 | 121 | 142 | 147 | 165 | 148 | 137 |
| 200 | 93 | 64 | 81 | 102 | 115 | 109 | 130 | 150 | 151 | 172 | 164 | 162 |
| 300 | 122 | 115 | 125 | 119 | 129 | 118 | 133 | 153 | 156 | 178 | 173 | 182 |
| 500 | 138 | 126 | 131 | 134 | 142 | 120 | 142 | 160 | 163 | 185 | 177 | 186 |
| 700 | 143 | 131 | 144 | 146 | 144 | 122 | 144 | 161 | 164 | 185 | 178 | 186 |
| 850 | 149 | 139 | 149 | 147 | 148 | 123 | 148 | 161 | 164 | 185 | 179 | 186 |

/Table 7.

Table 7. DIRECTION AND SPEED OF UPPER WINDS

ADEN 12°50'N. 45°01'E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mean Vector Wind (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 270 | 227 | 277 | 247 | 116 | 94 | 94 | 93 | 93 | 101 | Calm | 264 |
| kt. | 5 | 2 | 7 | 6 | 12 | 55 | 66 | 54 | 39 | 18 | | 7 |
| 150 ° | 293 | 257 | 291 | 271 | 147 | 99 | 93 | 93 | 100 | 124 | 275 | 273 |
| kt. | 15 | 10 | 19 | 18 | 3 | 49 | 71 | 60 | 46 | 15 | 14 | 23 |
| 200 ° | 279 | 241 | 291 | 273 | 290 | 96 | 90 | 91 | 98 | 125 | 281 | 273 |
| kt. | 17 | 11 | 25 | 19 | 2 | 36 | 48 | 45 | 36 | 13 | 14 | 26 |
| 300 ° | 289 | 270 | 296 | 268 | 279 | 88 | 80 | 87 | 94 | 115 | 269 | 262 |
| kt. | 15 | 7 | 18 | 16 | 6 | 21 | 23 | 24 | 20 | 8 | 8 | 17 |
| 500 ° | 59 | 77 | 166 | 62 | 66 | 48 | 45 | 51 | 53 | 85 | 99 | 192 |
| kt. | 3 | 1 | 1 | 1 | 9 | 16 | 13 | 13 | 10 | 12 | 3 | 3 |
| 700 ° | 34 | 56 | 101 | 56 | 64 | 53 | 29 | 49 | 52 | 35 | 43 | 39 |
| kt. | 3 | 4 | 3 | 8 | 9 | 7 | 3 | 4 | 7 | 15 | 9 | 5 |
| 850 ° | 63 | 64 | 86 | 87 | 99 | 236 | 278 | 267 | 258 | 80 | 68 | 63 |
| kt. | 7 | 6 | 5 | 10 | 1 | 2 | 5 | 5 | 1 | 7 | 13 | 11 |
| Standard Vector Deviation (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 18 | 16 | 17 | 15 | 19 | 22 | 20 | 18 | 18 | 16 | 15 | 21 |
| 150 | 22 | 23 | 21 | 21 | 26 | 21 | 14 | 22 | 21 | 19 | 20 | 27 |
| 200 | 25 | 25 | 29 | 23 | 27 | 19 | 15 | 16 | 18 | 17 | 23 | 31 |
| 300 | 19 | 20 | 22 | 20 | 21 | 12 | 12 | 8 | 11 | 13 | 19 | 23 |
| 500 | 13 | 12 | 10 | 13 | 14 | 13 | 11 | 10 | 11 | 16 | 13 | 15 |
| 700 | 6 | 11 | 10 | 12 | 13 | 11 | 9 | 10 | 11 | 10 | 13 | 11 |
| 850 | 9 | 10 | 11 | 12 | 8 | 9 | 10 | 10 | 8 | 10 | 7 | 9 |
| Mean Scalar Speed (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 16 | 14 | 15 | 14 | 18 | 55 | 66 | 54 | 40 | 21 | 13 | 18 |
| 150 | 23 | 21 | 24 | 23 | 22 | 50 | 71 | 61 | 47 | 21 | 20 | 29 |
| 200 | 26 | 23 | 31 | 26 | 23 | 37 | 50 | 46 | 37 | 19 | 22 | 36 |
| 300 | 22 | 18 | 24 | 22 | 19 | 22 | 25 | 24 | 22 | 14 | 18 | 25 |
| 500 | 12 | 10 | 8 | 12 | 14 | 19 | 16 | 16 | 13 | 15 | 12 | 13 |
| 700 | 10 | 10 | 9 | 13 | 14 | 11 | 9 | 9 | 11 | 17 | 14 | 11 |
| 850 | 10 | 10 | 9 | 13 | 7 | 8 | 10 | 10 | 7 | 10 | 15 | 14 |
| Number of Observations | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 63 | 57 | 66 | 78 | 87 | 70 | 72 | 84 | 114 | 130 | 122 | 122 |
| 150 | 80 | 71 | 84 | 85 | 100 | 94 | 93 | 120 | 130 | 142 | 140 | 144 |
| 200 | 84 | 74 | 85 | 86 | 103 | 95 | 96 | 124 | 134 | 145 | 140 | 147 |
| 300 | 86 | 79 | 85 | 88 | 104 | 100 | 100 | 126 | 141 | 150 | 141 | 150 |
| 500 | 87 | 80 | 88 | 89 | 106 | 100 | 103 | 128 | 143 | 150 | 144 | 151 |
| 700 | 87 | 80 | 88 | 89 | 106 | 102 | 103 | 128 | 143 | 150 | 144 | 151 |
| 850 | 87 | 80 | 88 | 89 | 106 | 102 | 103 | 128 | 142 | 150 | 144 | 151 |

/Table 8.

## Table 8. DIRECTION AND SPEED OF UPPER WINDS

## NAIROBI 01°17'S. 36°50'E. 1948-50

| Level | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mean Vector Wind (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 ° | 87 | 89 | 83 | 69 | 128 | 88 | 86 | 98 | 85 | 98 | 345 | 340 |
| kt. | 11 | 13 | 9 | 2 | 4 | 15 | 14 | 10 | 10 | 4 | 4 | 5 |
| 150 ° | 88 | 102 | 101 | 261 | 56 | 96 | 86 | 97 | 96 | 226 | 291 | 226 |
| kt. | 13 | 16 | 13 | 4 | 5 | 10 | 30 | 18 | 9 | 5 | 14 | 1 |
| 200 ° | 109 | 110 | 109 | 156 | 324 | 107 | 101 | 95 | 110 | 233 | 286 | 139 |
| kt. | 23 | 23 | 17 | 1 | 5 | 5 | 13 | 11 | 5 | 5 | 11 | 11 |
| 300 ° | 83 | 94 | 91 | 97 | 234 | 172 | 208 | 124 | 121 | 128 | 45 | 78 |
| kt. | 15 | 15 | 15 | 7 | 2 | 3 | 3 | 4 | 3 | 5 | 2 | 8 |
| 500 ° | 88 | 85 | 85 | 86 | 71 | 34 | 31 | 21 | 67 | 88 | 87 | 97 |
| kt. | 13 | 14 | 18 | 13 | 6 | 4 | 5 | 6 | 5 | 11 | 16 | 10 |
| 700 ° | 36 | 40 | 54 | 72 | 104 | 115 | 260 | 211 | 74 | 90 | 74 | 51 |
| kt. | 12 | 15 | 17 | 11 | 8 | 3 | 3 | 2 | 3 | 10 | 13 | 13 |
| Standard Vector Deviation (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 20 | 20 | 18 | 16 | 15 | 22 | 22 | 19 | 18 | 17 | 21 | 20 |
| 150 | 22 | 24 | 21 | 17 | 22 | 25 | 28 | 24 | 26 | 16 | 19 | 27 |
| 200 | 23 | 23 | 18 | 17 | 18 | 18 | 22 | 18 | 21 | 15 | 16 | 23 |
| 300 | 17 | 15 | 16 | 12 | 15 | 14 | 17 | 13 | 15 | 11 | 14 | 17 |
| 500 | 11 | 13 | 10 | 11 | 10 | 11 | 11 | 10 | 9 | 8 | 7 | 11 |
| 700 | 10 | 11 | 9 | 8 | 9 | 10 | 9 | 9 | 9 | 7 | 8 | 10 |
| Mean Scalar Speed (knots) | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 21 | 20 | 18 | 16 | 14 | 21 | 21 | 17 | 18 | 15 | 19 | 18 |
| 150 | 22 | 24 | 21 | 15 | 19 | 22 | 35 | 25 | 23 | 15 | 21 | 23 |
| 200 | 28 | 28 | 22 | 15 | 17 | 15 | 21 | 20 | 20 | 14 | 16 | 21 |
| 300 | 21 | 19 | 20 | 12 | 12 | 13 | 16 | 12 | 13 | 11 | 13 | 17 |
| 500 | 15 | 17 | 20 | 14 | 11 | 10 | 10 | 10 | 10 | 13 | 17 | 14 |
| 700 | 14 | 17 | 18 | 12 | 11 | 9 | 9 | 8 | 9 | 12 | 14 | 16 |
| Number of Observations | | | | | | | | | | | | |
| mb. | | | | | | | | | | | | |
| 100 | 49 | 46 | 48 | 58 | 24 | 41 | 55 | 52 | 65 | 80 | 77 | 67 |
| 150 | 61 | 50 | 57 | 63 | 30 | 46 | 57 | 58 | 69 | 84 | 80 | 74 |
| 200 | 61 | 51 | 58 | 64 | 35 | 48 | 59 | 59 | 72 | 85 | 81 | 75 |
| 300 | 62 | 52 | 59 | 64 | 38 | 48 | 59 | 64 | 74 | 85 | 84 | 75 |
| 500 | 66 | 54 | 61 | 65 | 42 | 49 | 61 | 68 | 76 | 88 | 88 | 76 |
| 700 | 65 | 52 | 61 | 64 | 44 | 49 | 61 | 68 | 75 | 86 | 89 | 77 |

# VECTOR RESULTANT WINDS AT GIBRALTAR

1948-50

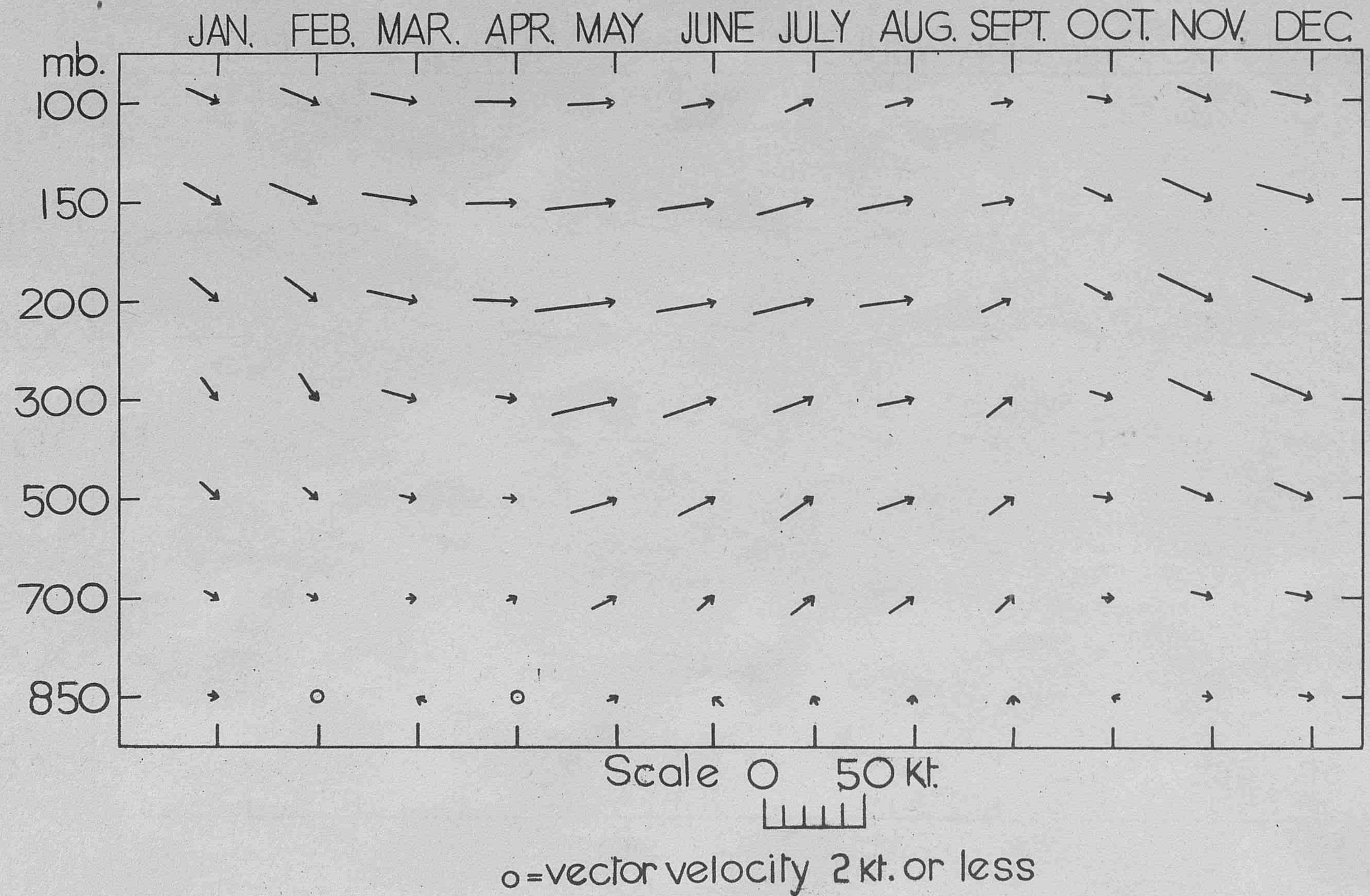


Fig. 1(a).

VECTOR RESULTANT WINDS AT GIBRALTAR

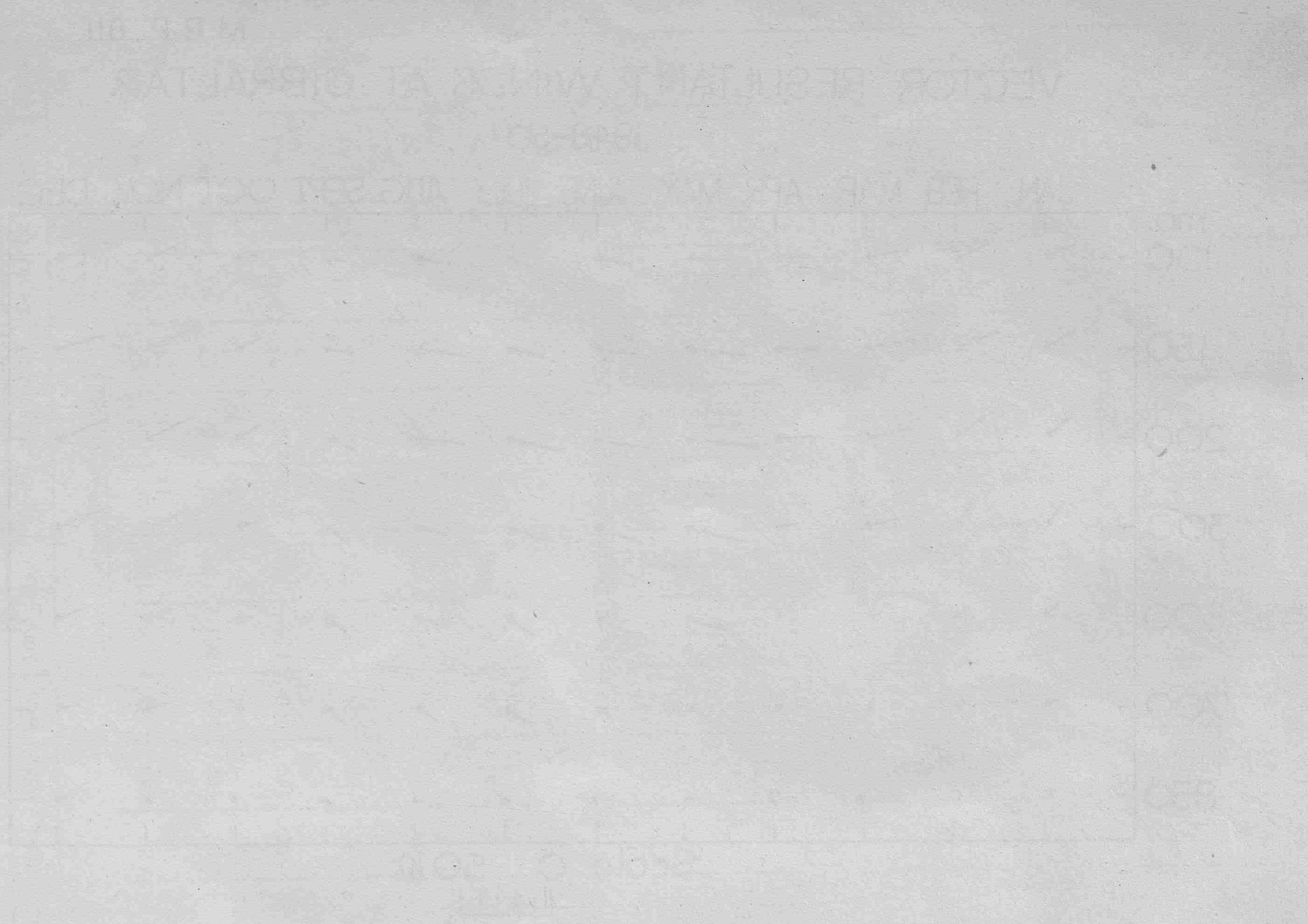


Fig. 104

# VECTOR RESULTANT WINDS AT MALTA

## 1948-50

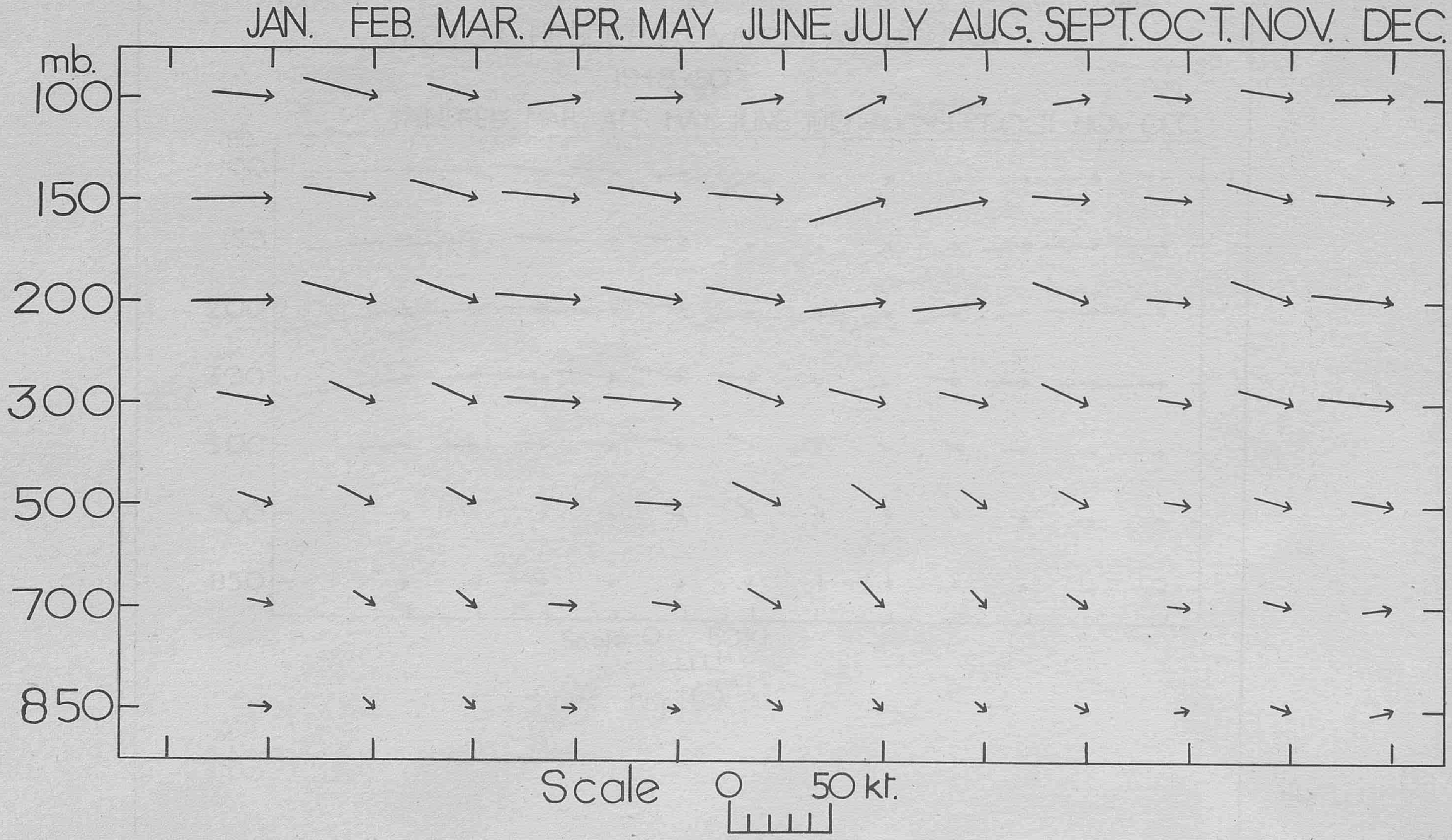


Fig. 1(b).

# VECTOR RESULTANT WINDS AT BENINA

1948-50

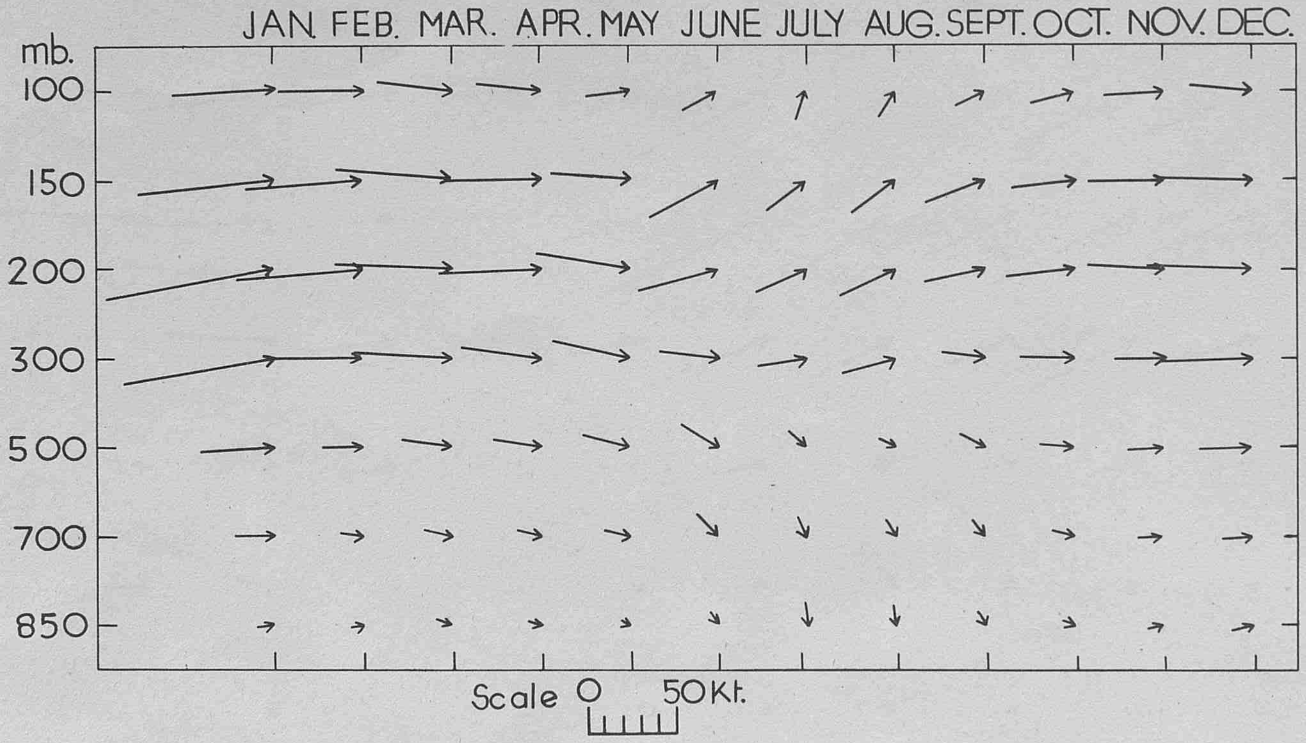
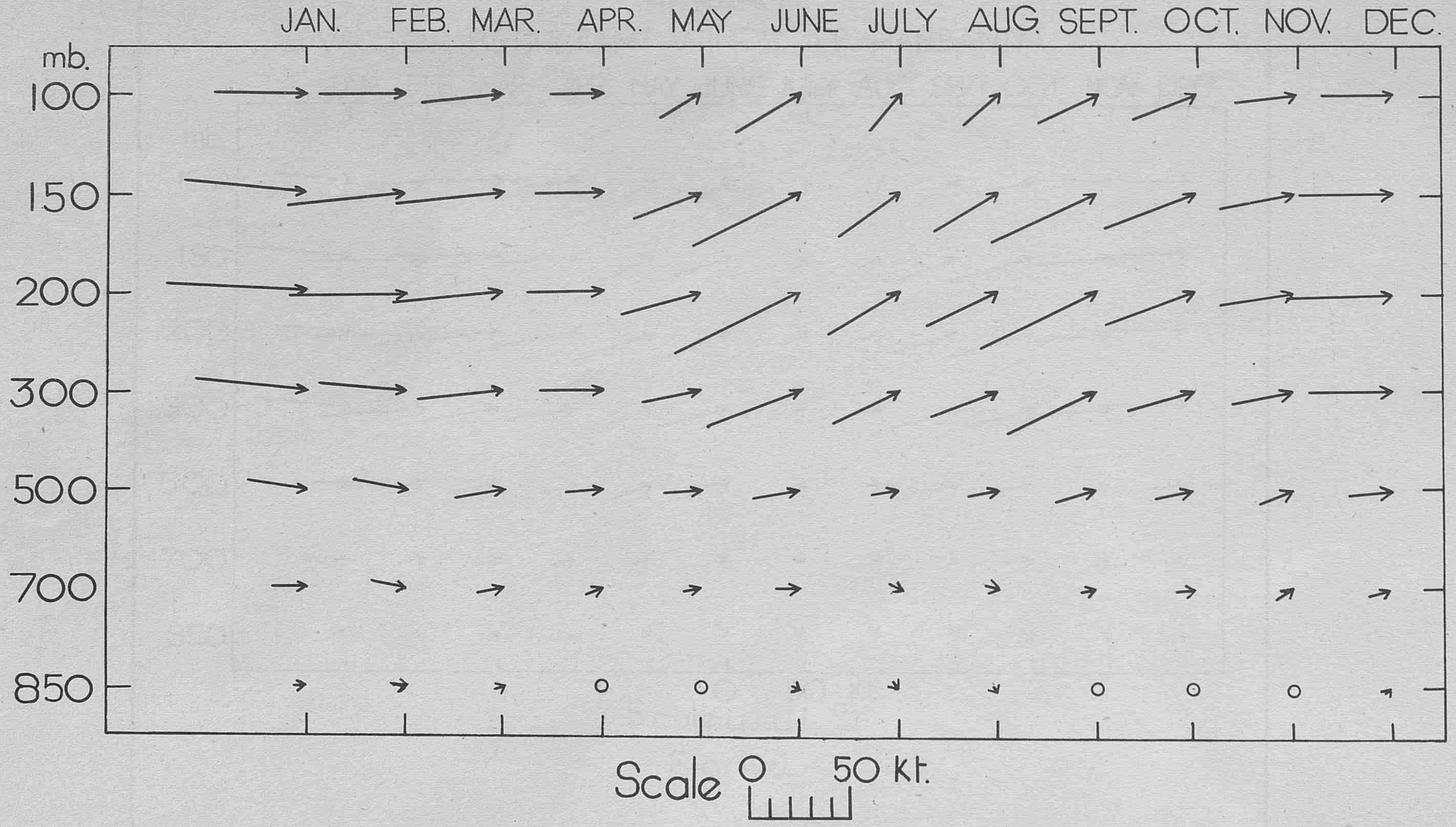


Fig. 1 (c).

# VECTOR RESULTANT WINDS AT NICOSIA

## 1948-50



o = vector velocity 2 kt. or less

Fig. 1(d).

## 1948-50
## VECTOR RESULTANT WINDS AT HABBANIYA

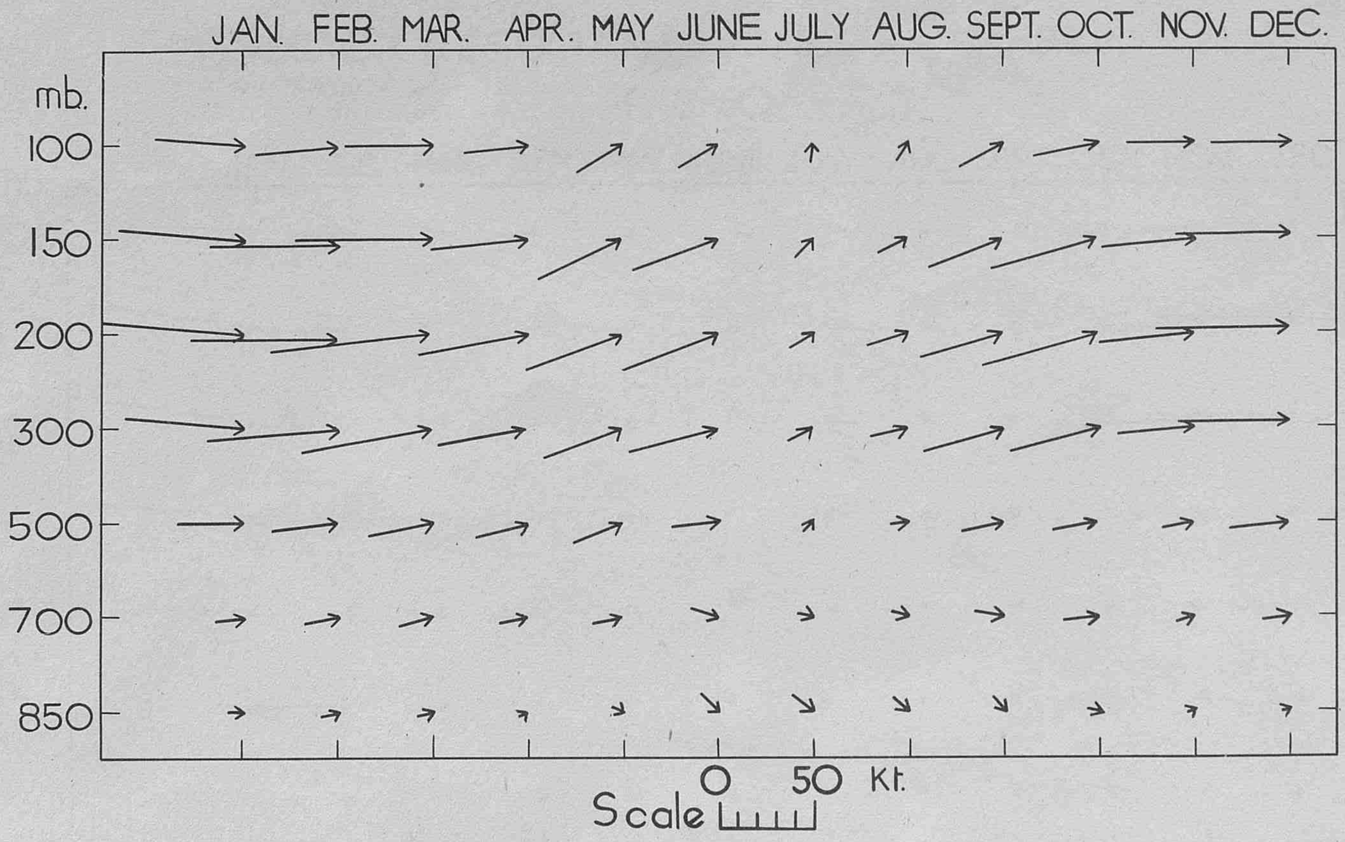
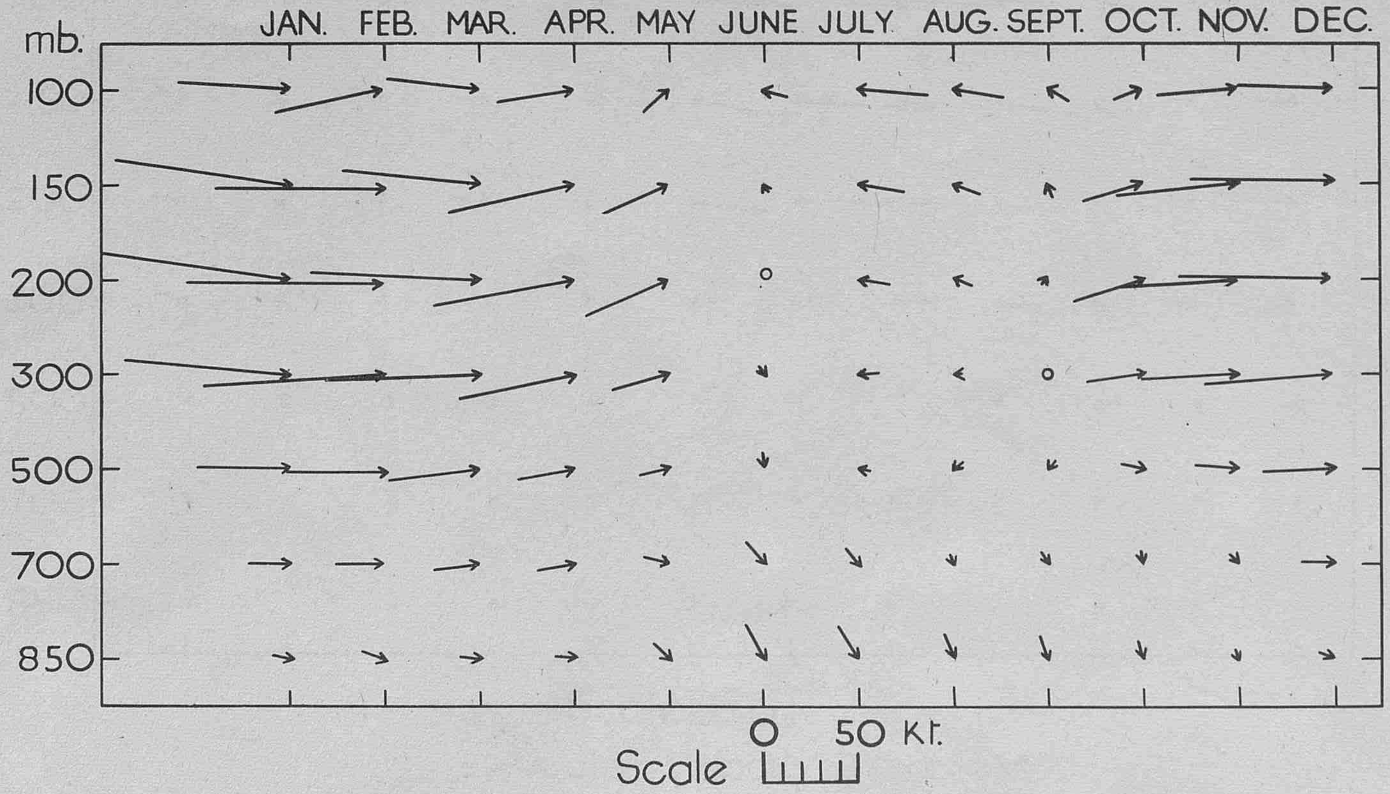


Fig. 1(e).

## VECTOR RESULTANT WINDS AT BAHREIN
## 1948-50



o = vector velocity 2 kt. or less

Fig. 1(F).

M.R.P. 811

# VECTOR RESULTANT WINDS AT ADEN

1948-50

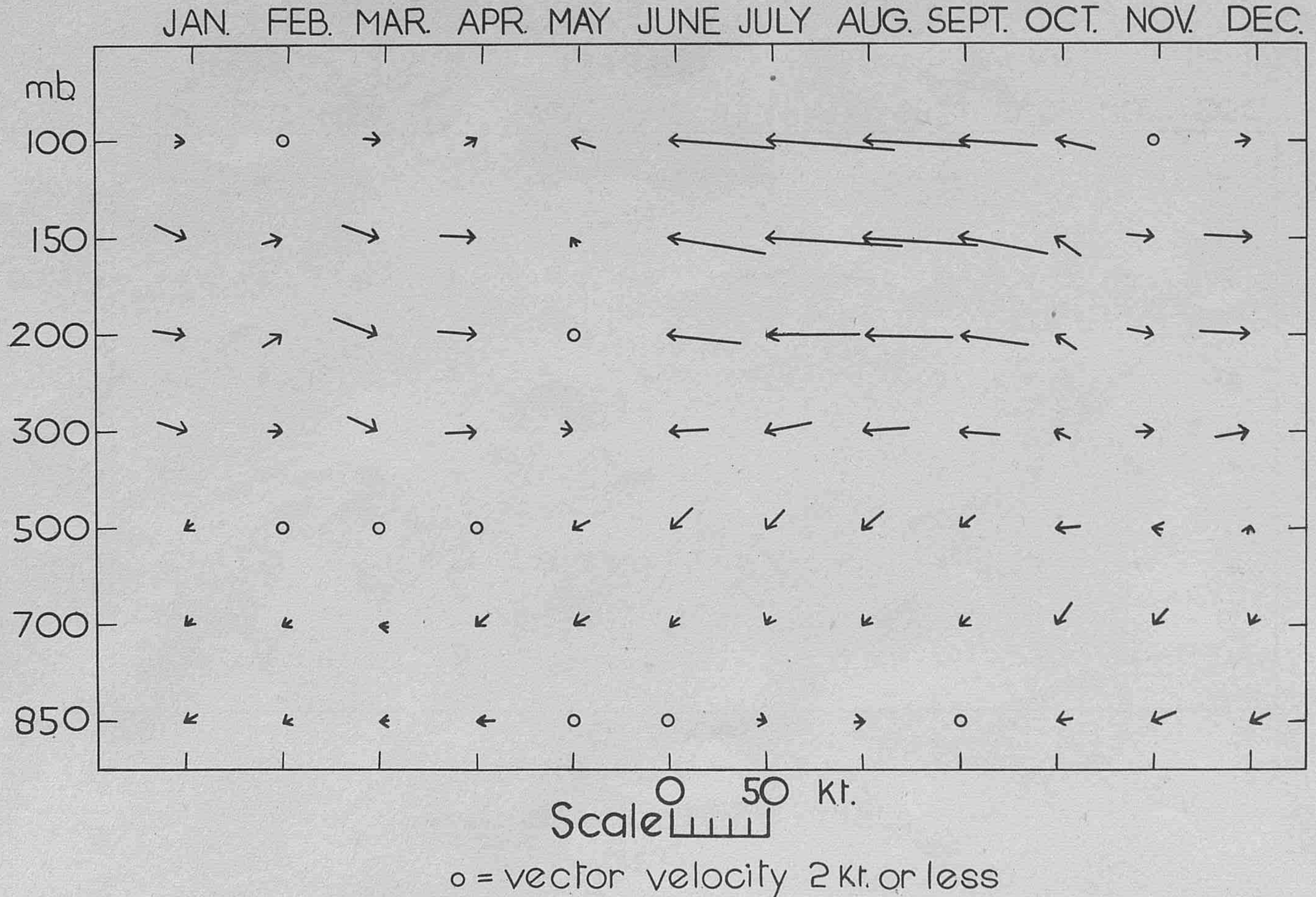
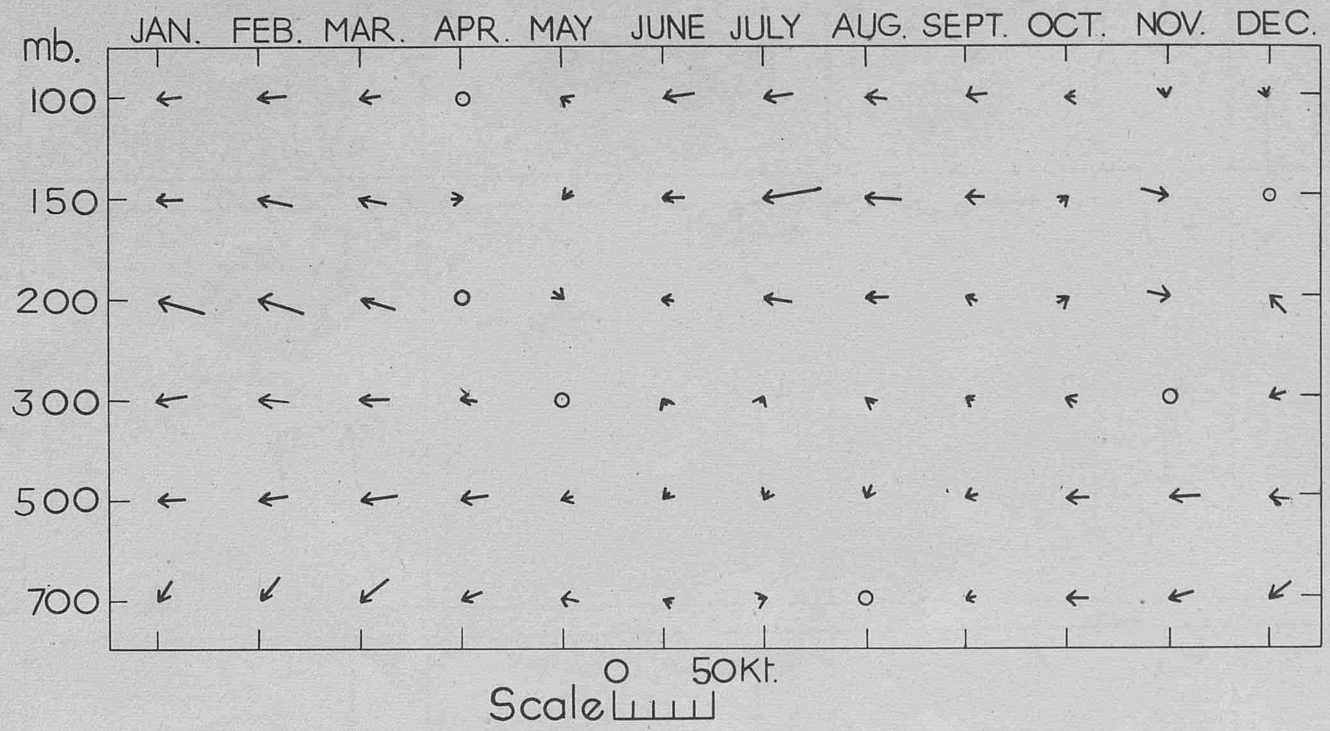


Fig. 1(g).

## VECTOR RESULTANT WINDS AT NAIROBI

1948-50



o=vector velocity 2kt. or less

Fig. 1(h).

# MEAN VECTOR WINDS, 1948-50

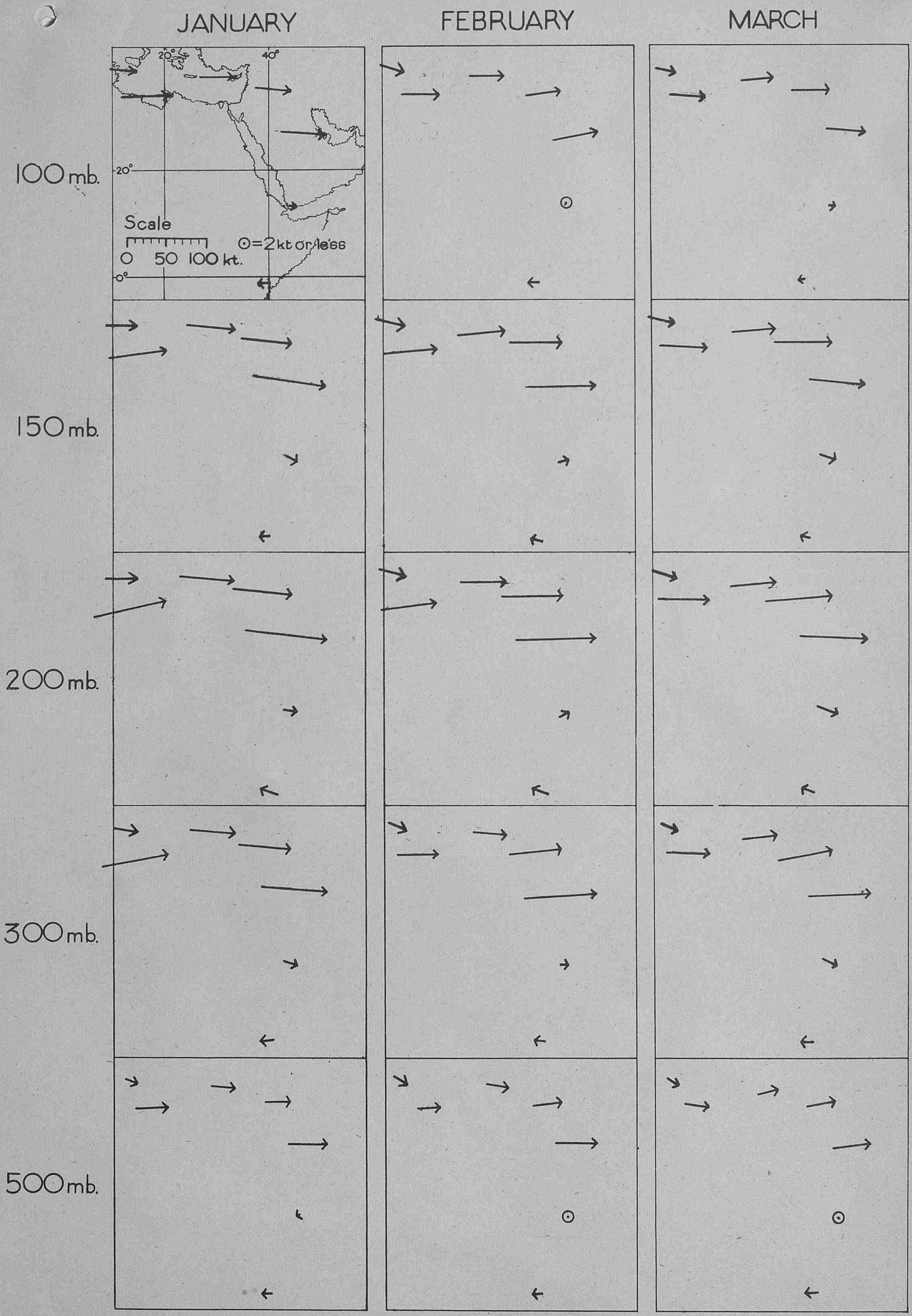

JANUARY
FEBRUARY
MARCH
100 mb.
150 mb.
200 mb.
300 mb.
500 mb.
20°
40°
20°
0°
Scale
0 50 100 kt.
⊙=2kt or less

Fig. 2 (a).

# MEAN VECTOR WINDS, 1948-50

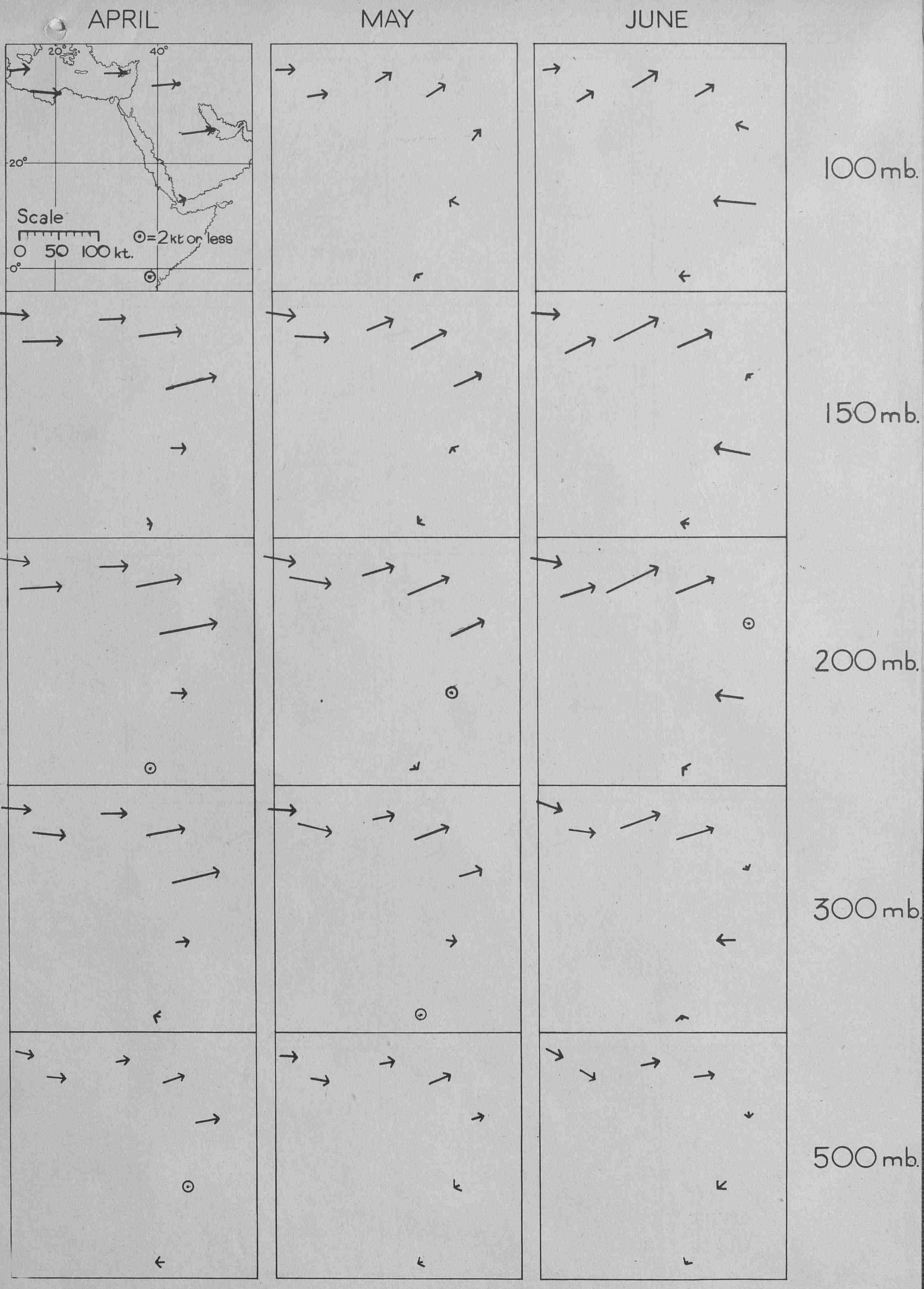

APRIL
MAY
JUNE
20°
40°
20°
0°
Scale
0 50 100 kt.
⊙=2 kt or less
100 mb.
150 mb.
200 mb.
300 mb.
500 mb.

Fig. 2 (b).

# MEAN VECTOR WINDS, 1948-50

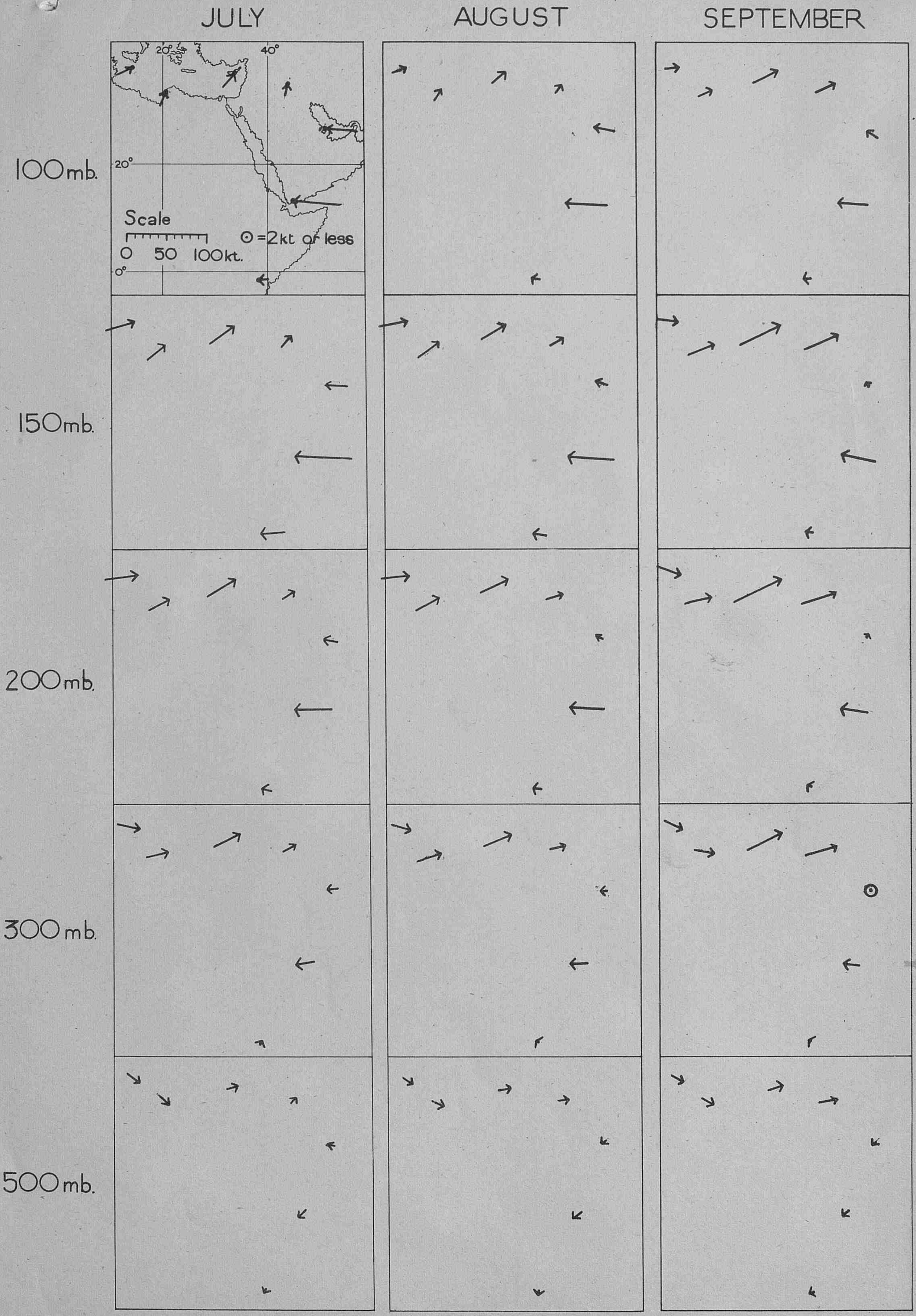

JULY
AUGUST
SEPTEMBER
100mb.
150mb.
200mb.
300mb.
500mb.
20°
40°
20°
0°
Scale
0 50 100kt.
⊙ =2kt or less

Fig. 2(c).

# MEAN VECTOR WINDS, 1948-50

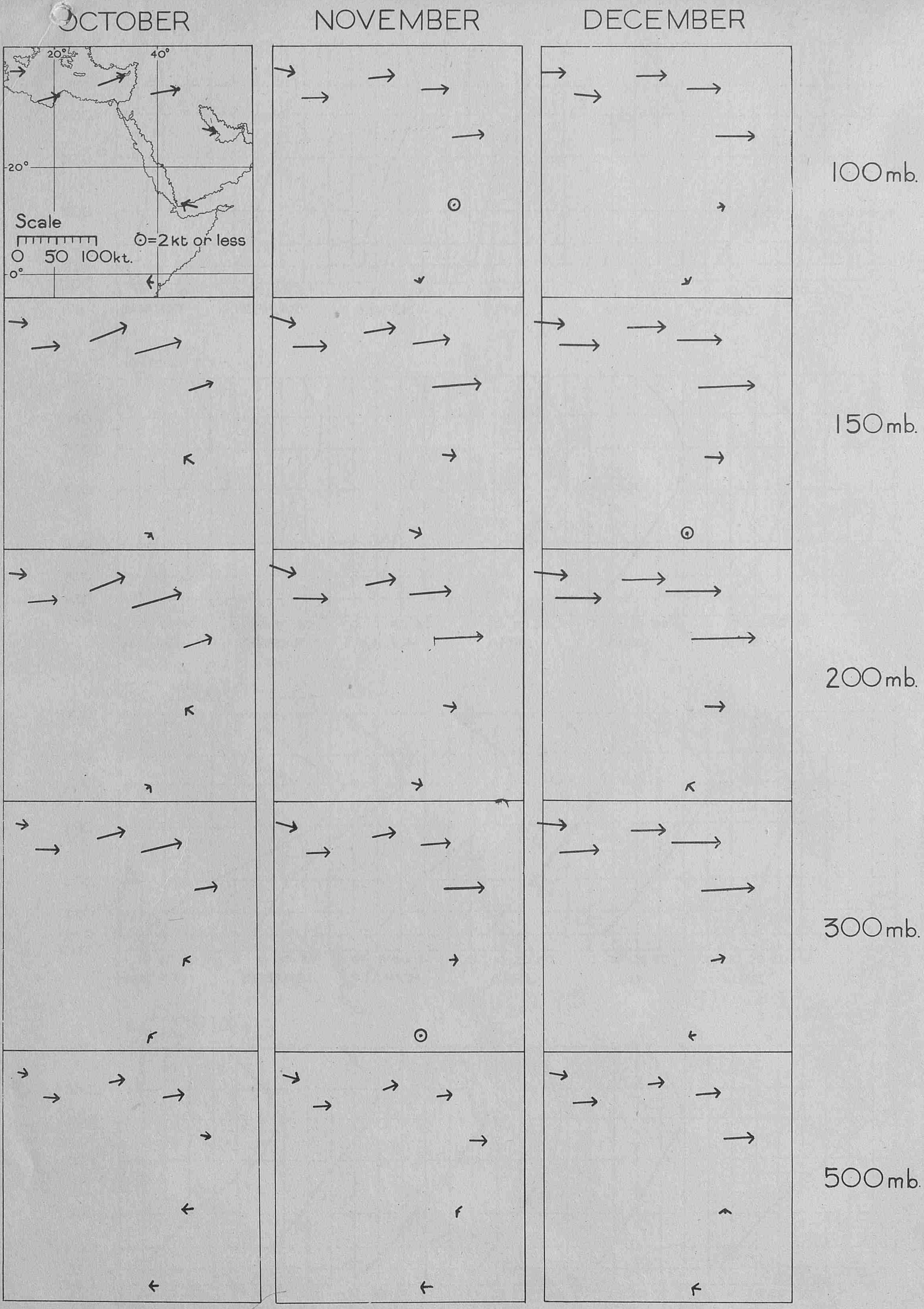

OCTOBER
NOVEMBER
DECEMBER
20°
40°
20°
0°
Scale
0 50 100kt.
⊙=2kt or less
100 mb.
150 mb.
200 mb.
300 mb.
500 mb.

Fig. 2(d).

# VARIATION OF WIND WITH HEIGHT

Mean vector wind

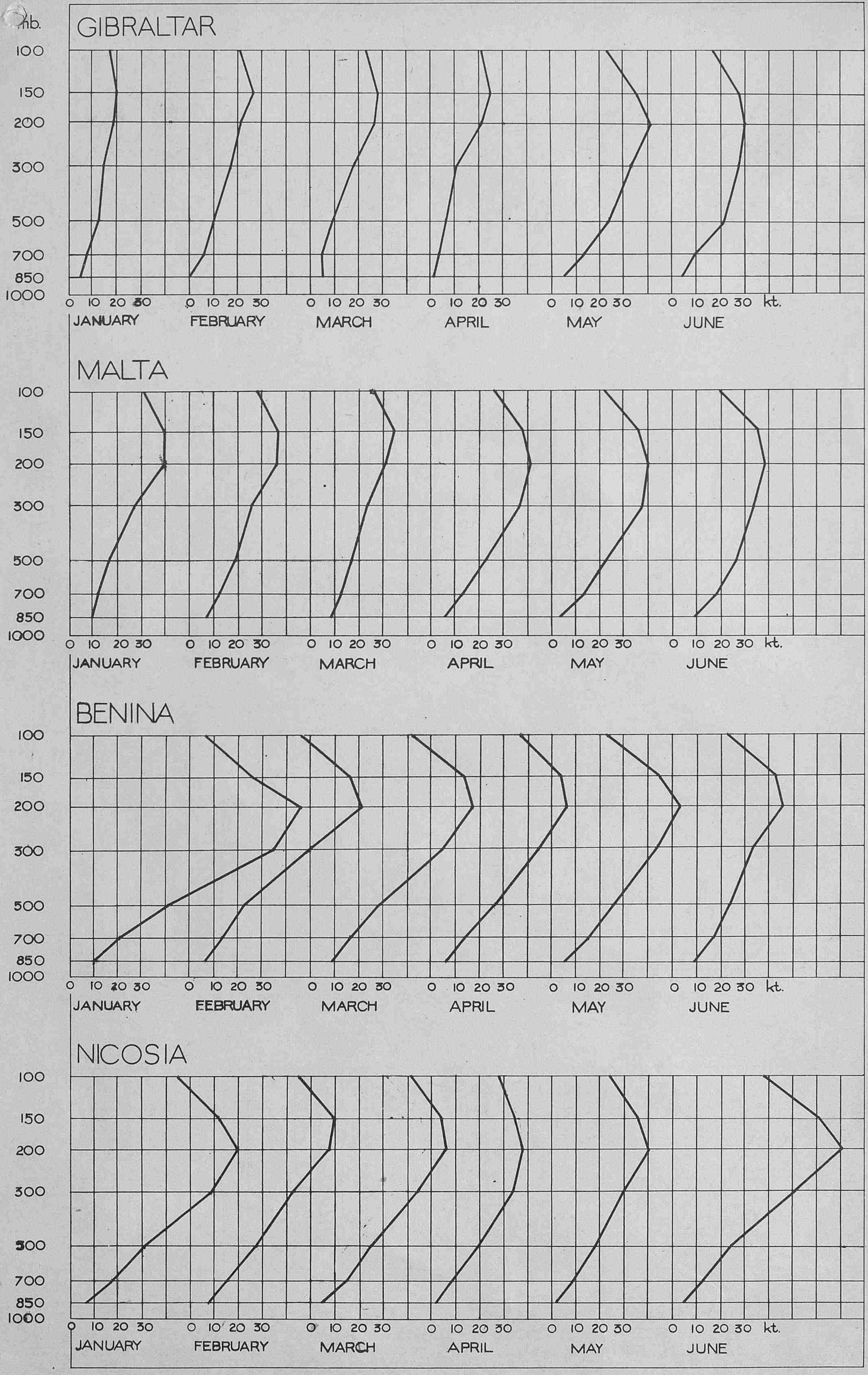


Fig. 3(a).

# VARIATION OF WIND WITH HEIGHT

Mean vector wind

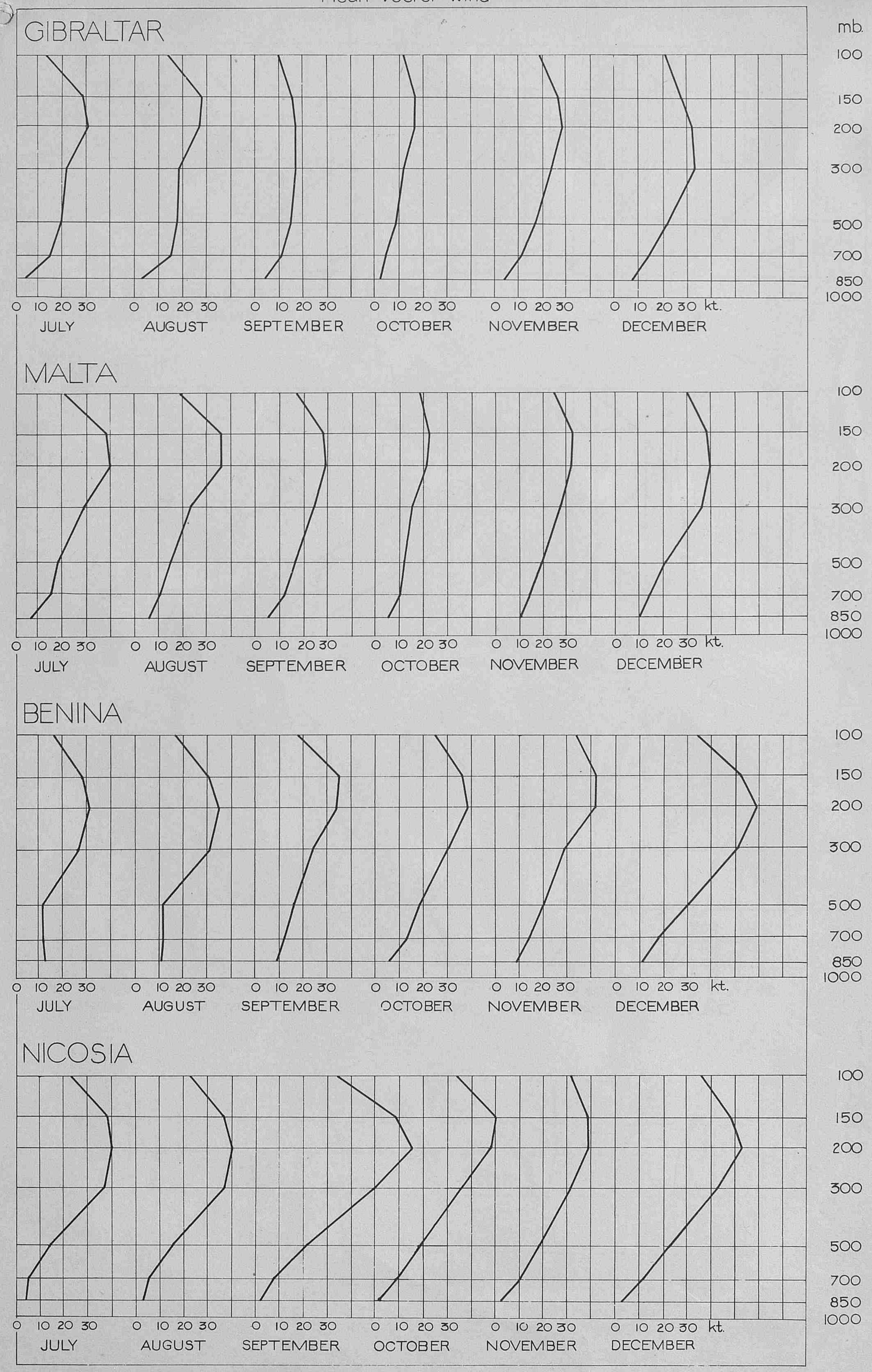


Fig. 3(b).

VARIATION OF WIND WITH HEIGHT
Mean vector wind

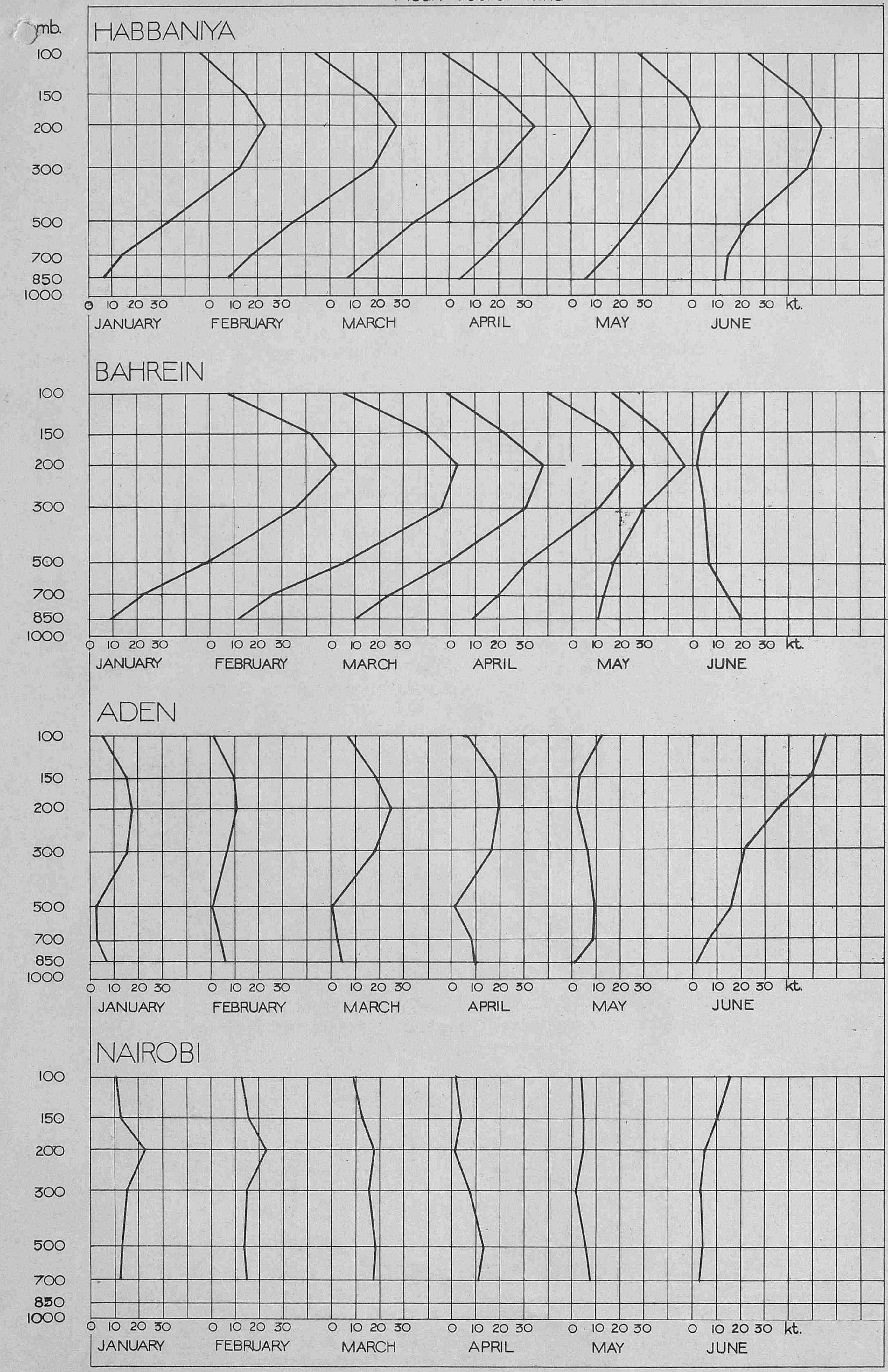


Fig. 3(c).

VARIATION OF WIND WITH HEIGHT

Mean vector wind

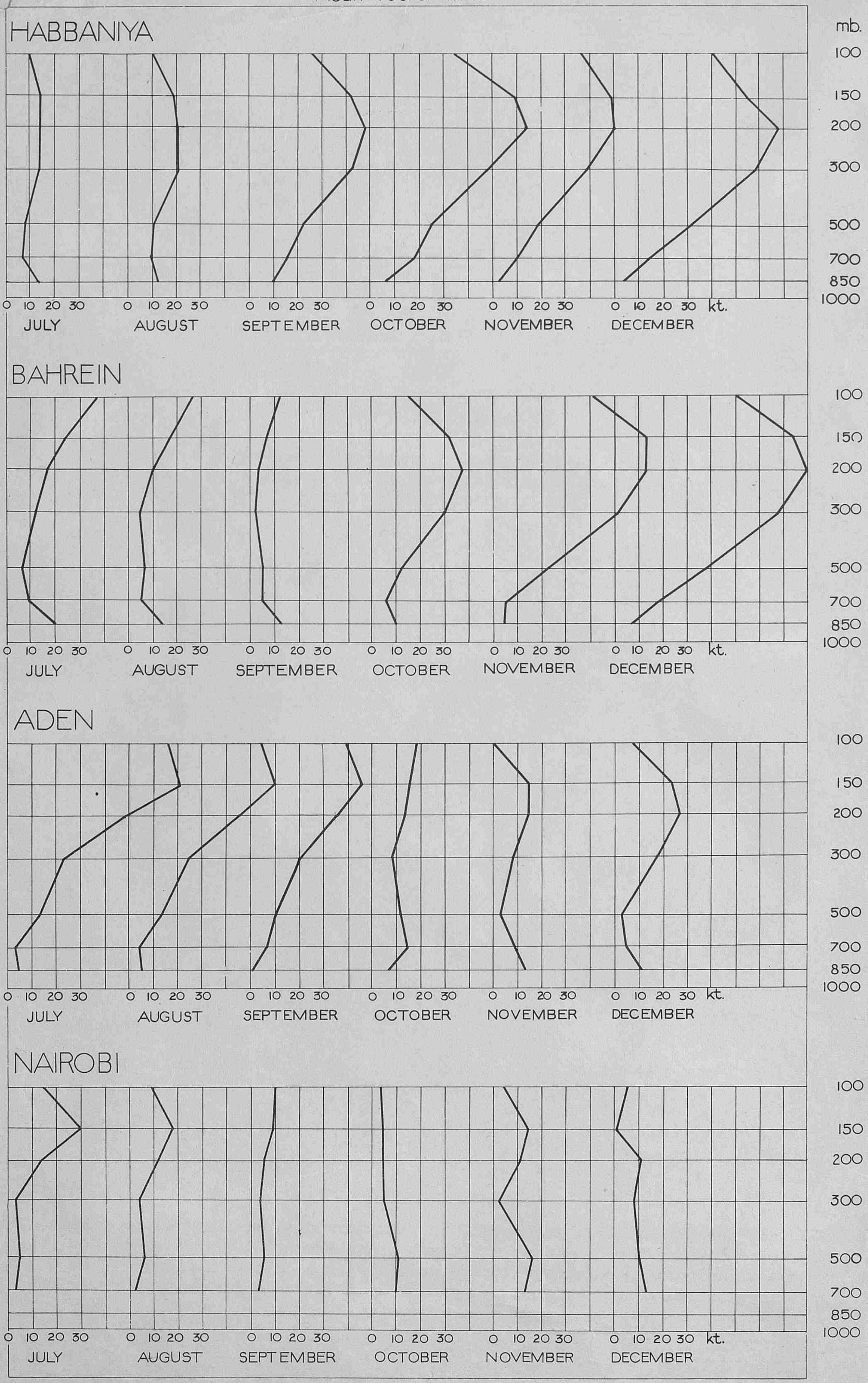


Fig. 3(d).

Control Number
[redacted]

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A copy is available in the Library of the Meteorological Office."

M.R.P. 812
S.C.II/103
20th May, 1953

AIR WEATHER SERVICE
TECHNICAL REFERENCE FILES

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TECHNICAL REFERENCE FILES

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# HIGH ALTITUDE WINDS AT O.W.S. "POLAR FRONT"

by

R. R. Fotheringham

1. Microfilms of upper air data from the Ocean Weather Ship on station 'M' (66°N 02°E) have been examined and statistics relating to upper wind velocities have been extracted for the period June 1948-April 1952.

## Wind Speed

2. A table was compiled setting out the number of days in each month throughout the whole period when winds of 80 knots or more were observed, within certain intervals of height. The table (available in manuscript) is not reproduced here but Figures 1 - 4 have been drawn from these data to show the variations of frequency of occurrence of strong winds during the four seasons. (The rather odd intervals of height in feet result from conversion of the original heights in metres). The obvious features revealed by these graphs are

(a) the marked maximum frequency at all seasons between about 25,000 feet and 35,000 feet;

(b) the notable decrease in summer as compared with the other seasons which show little variation.

Figure 5 illustrates (b) above for the height interval corresponding to maximum frequency.

3. It is interesting to note that in M.R.P. 761 D.H. Johnson shows that the mean cross-section from Greenland to Tunis for January 1950 has a maximum in the upper flow centred just below 300 mb and located about 61°N. O.W.S. 'M' in 66°N lies more or less downwind from this position assuming the mean flow is South Westerly and reference to Figure 4 shows that the maximum frequency of strong winds occurs at about this height.

4. The number of days during the six months October to March for the four winters when the wind exceeded 160 knots is shown in Table I. The frequencies are grouped according to height; maximum frequency of these very strong winds again occurs around 30,000 feet. The absolute maximum wind during the period was observed on 6th January 1952; it was 210 knots from 260° at a height of 28,500 feet

## Wind Direction

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Committee (London),
M.R.P. No. 812 by Instr/Lt.
R.R. Fotheringham
A copy is available in the Library of
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M.R.P. 812
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### Wind Direction

5. The mean seasonal distributions of the directions of winds greater than 80 knots and greater than 100 knots are shown as percentages in Figures 6 - 9. As might be expected winds with a westerly component predominate although in summer there is an appreciable proportion of strong winds from a South Easterly direction.

/General Remarks

DSS
T.F.
(W1-V1-4+Dc)

## General Remarks

6. It is plain from the detailed results that there is considerable variation in frequency from year to year and therefore means derived from so short a period are not necessarily representative.

At high levels the number of observations is reduced due to the larger number of ascents which terminate before reaching these levels; the reliability of statistics is thereby affected.

## Comparison with Larkhill and Lerwick

7. It was thought worth while to compare results from MIKE with data for Larkhill and Lerwick published by the Meteorological Office. For this purpose the frequency of observation, rather than number of days per month, of strong winds of particular speeds was used. Figures 10 - 12 set out the frequencies for the levels 400 mb, 300 mb and 200 mb of all winds greater than 80 knots.

Over the eastern North Atlantic during the period examined there is no evidence of a seasonal change in the latitude of the maximum frequency of strong winds: in fact, frequencies at MIKE are consistently lower than at Larkhill and Lerwick.

## Acknowledgment

Thanks are due to the Director, Meteorological Office, for loan of the microfilms.

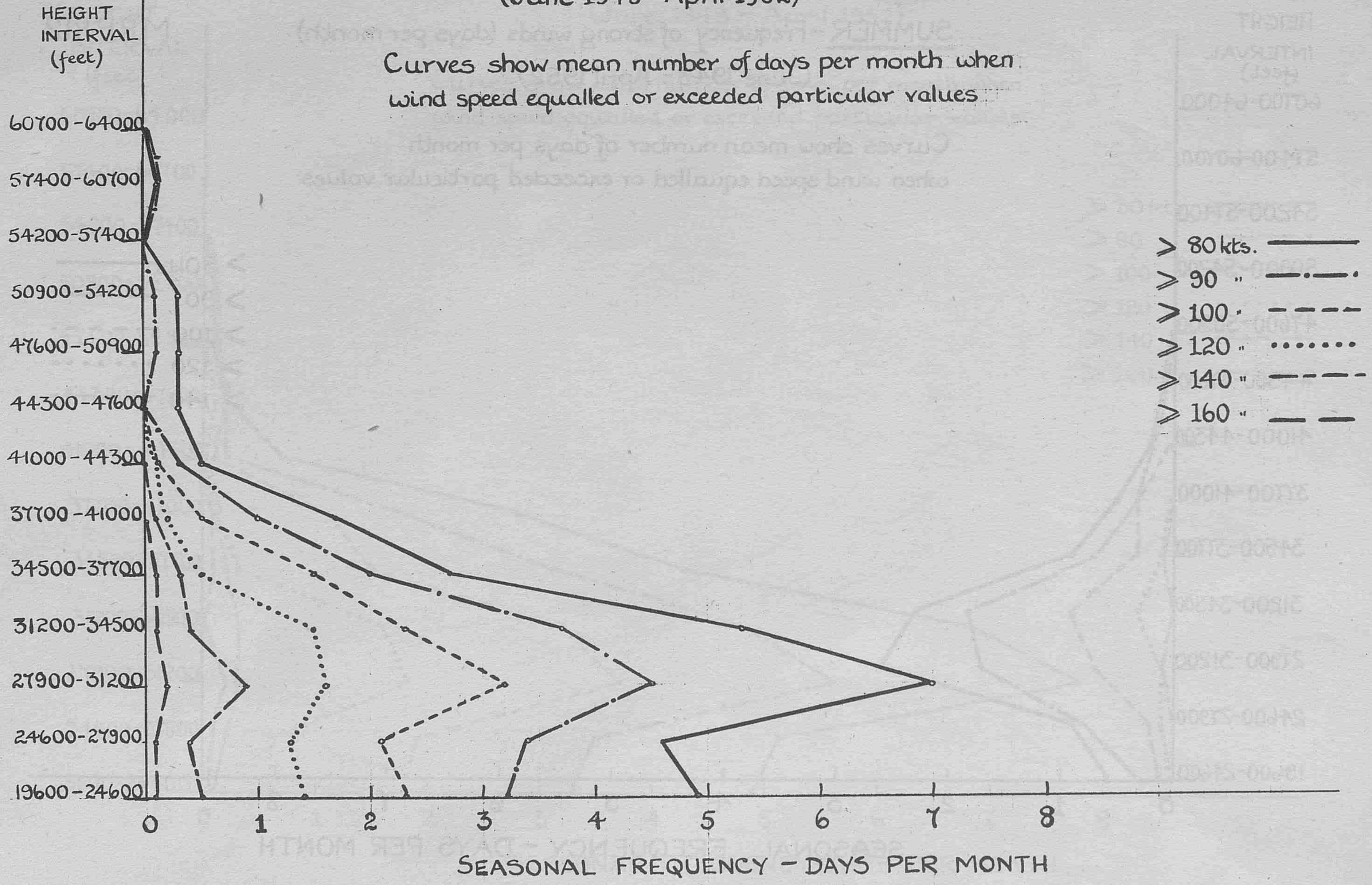
TABLE I

Number of days during the period October - March 1948 - 1952 when the wind exceeded 160 knots

| HEIGHT | 160 - 169 | 170 - 179 | 180 - 189 | 190 - 199 | 200 - 209 | 210 - 219 |
|---|---|---|---|---|---|---|
| A<br>19,600-24,600 | 2 | 1 | - | - | - | - |
| B<br>24,600-27,900 | 2 | 1 | - | - | - | - |
| C<br>27,900-31,200 | 3 | 3 | 2 | - | - | 1 |
| D<br>31,200-34,500 | 3 | 1 | - | - | - | - |
| E<br>34,500-37,700 | 1 | 1 | - | - | - | - |
| F<br>37,700-41,000 | 1 | - | - | - | - | - |

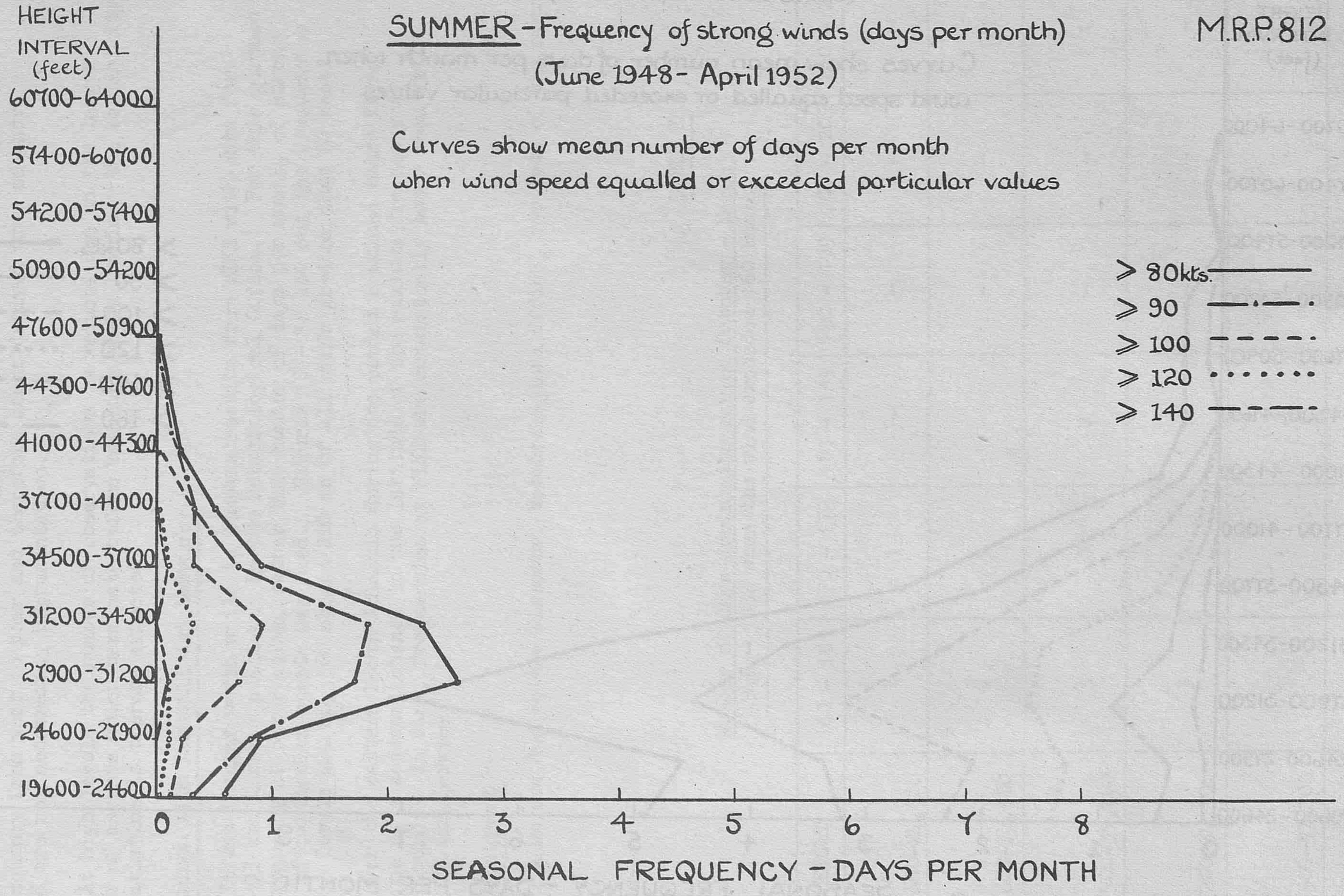
## SPRING - Frequency of strong winds (days per month)

(June 1948 - April 1952)



Curves show mean number of days per month when wind speed equalled or exceeded particular values

FIG. 1



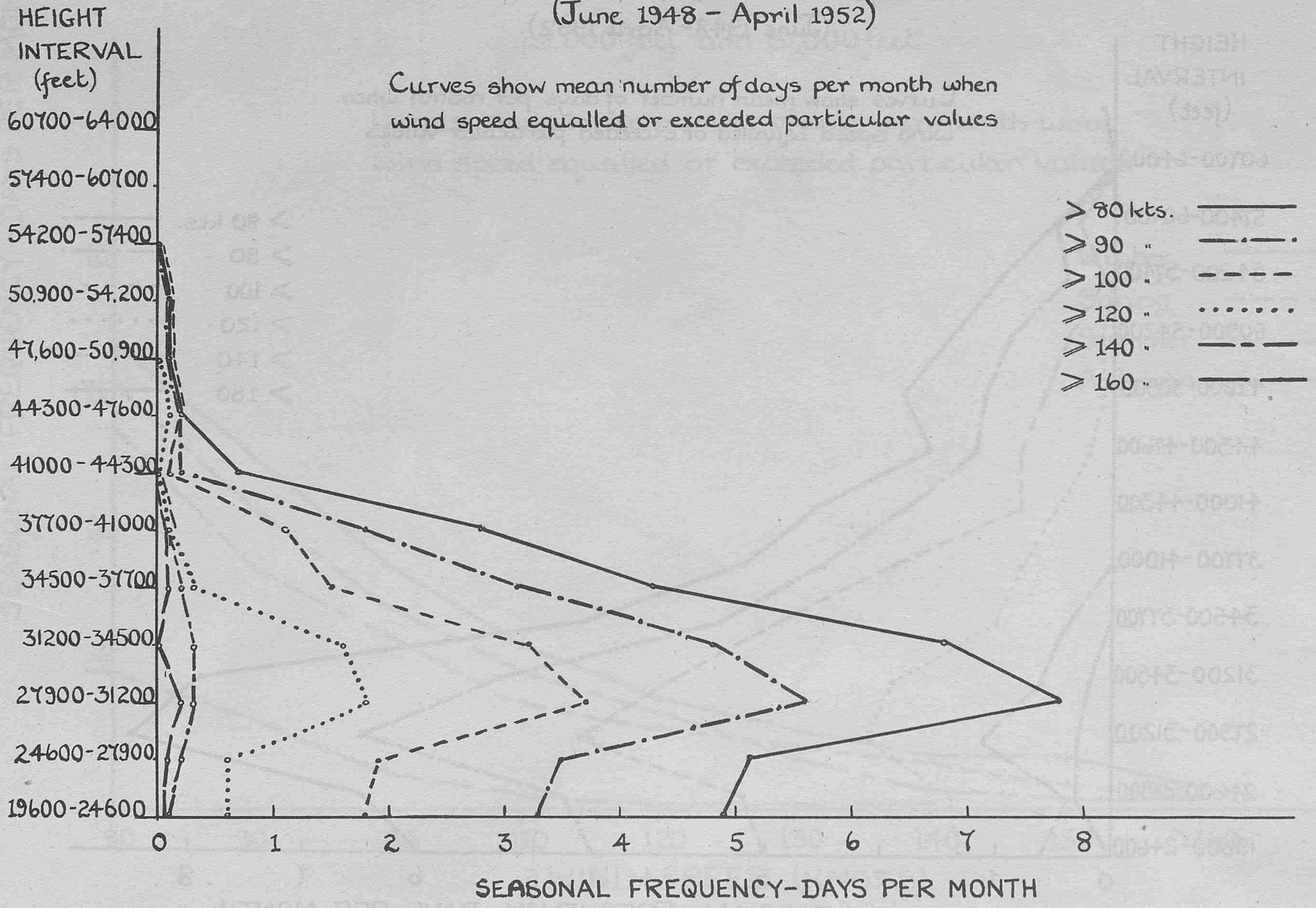
**SUMMER** - Frequency of strong winds (days per month)
(June 1948 - April 1952)

Curves show mean number of days per month when wind speed equalled or exceeded particular values

FIG. 2

## AUTUMN - Frequency of strong winds (days per month)

(June 1948 - April 1952)



Curves show mean number of days per month when wind speed equalled or exceeded particular values

FIG. 3

WINTER - Frequency of strong winds (days per month)
(June 1948 - April 1952)

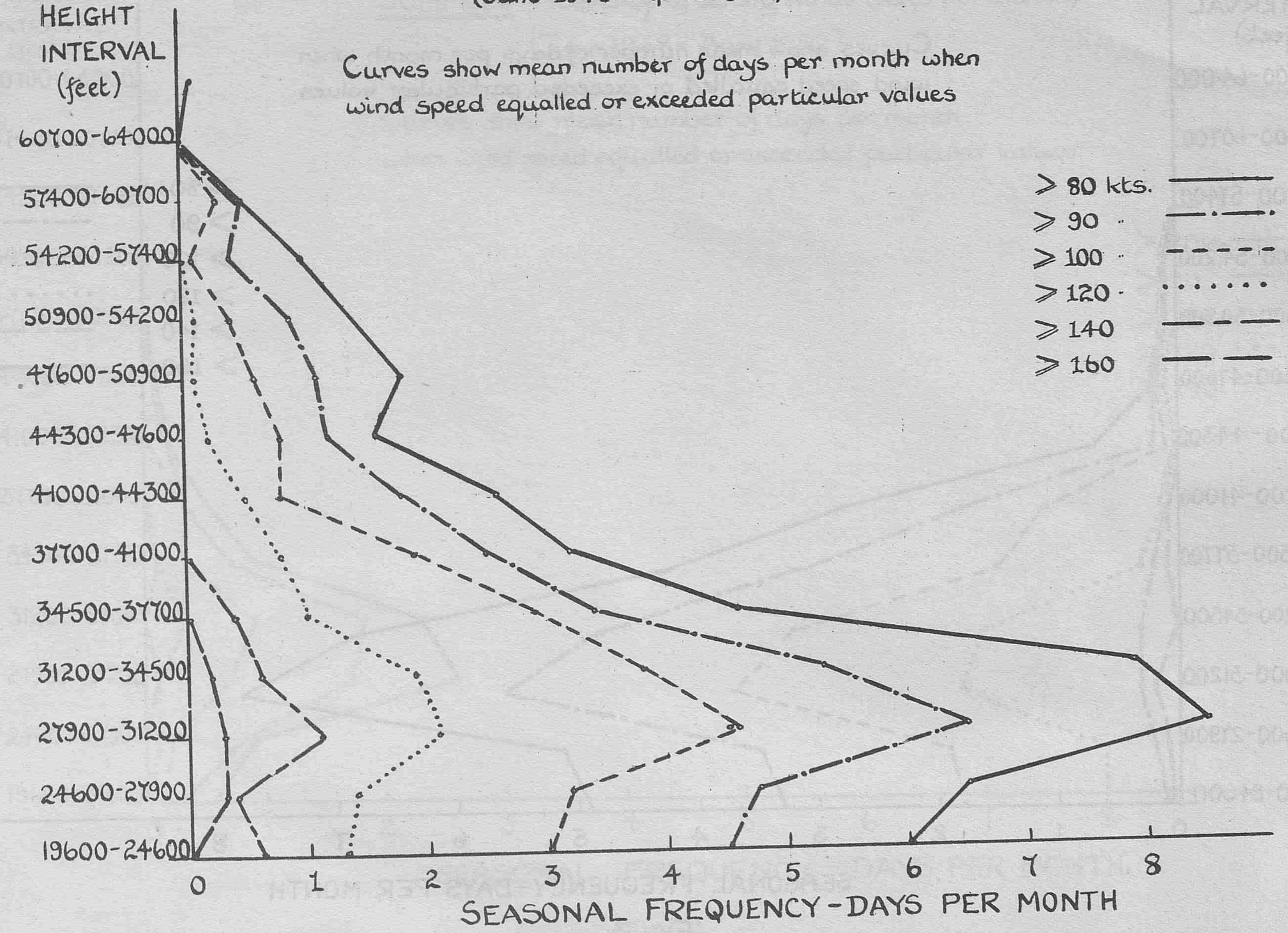


FIG. 4

Winds over O.W.S. "Polar Front" (66°N. 02°E.) between 28,000 feet and 31,000 feet

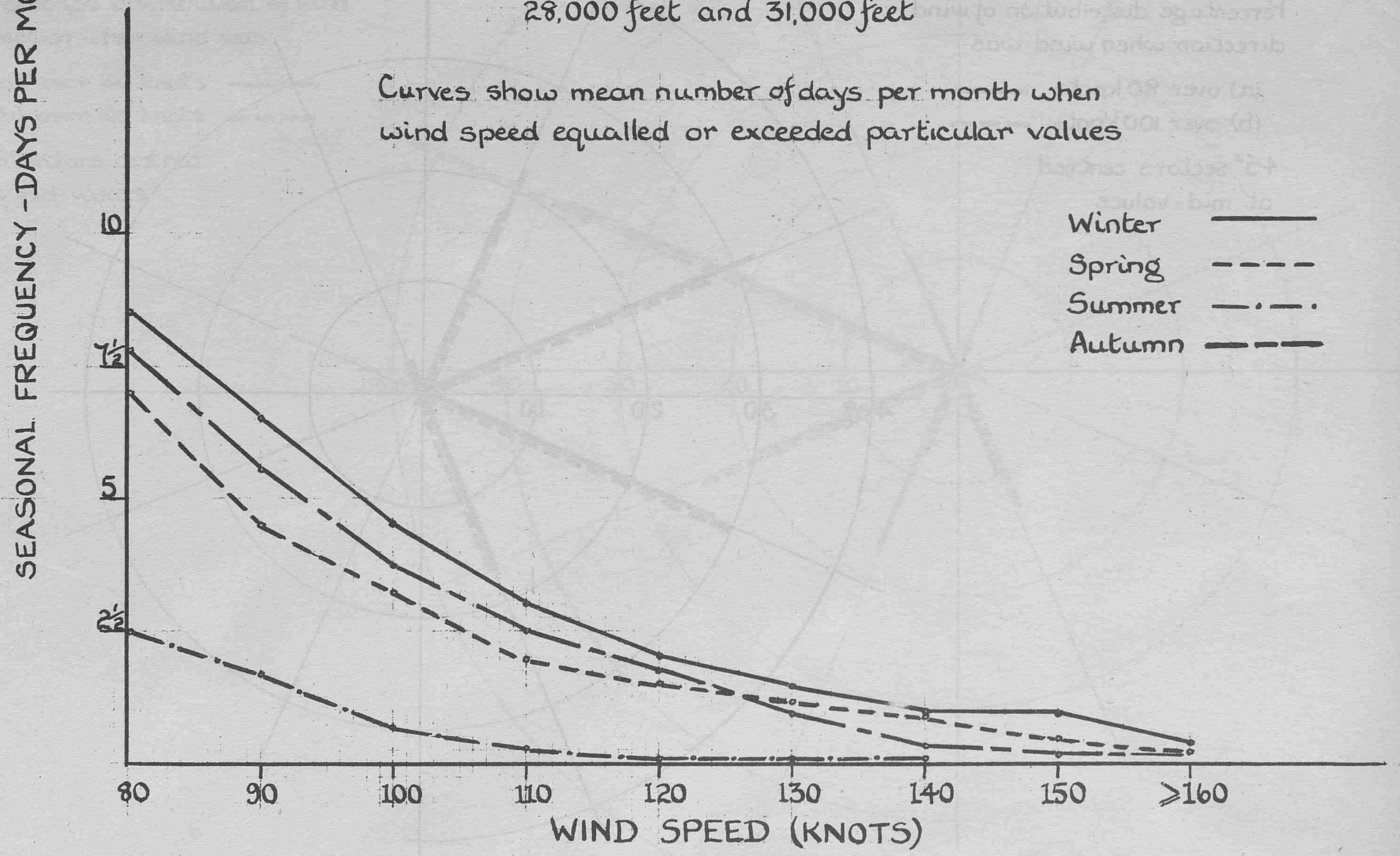


FIG. 5

## SPRING

Percentage distribution of wind direction when wind was

(a) over 80 knots ———

(b) over 100 knots - - -

45° sectors centred at mid-values

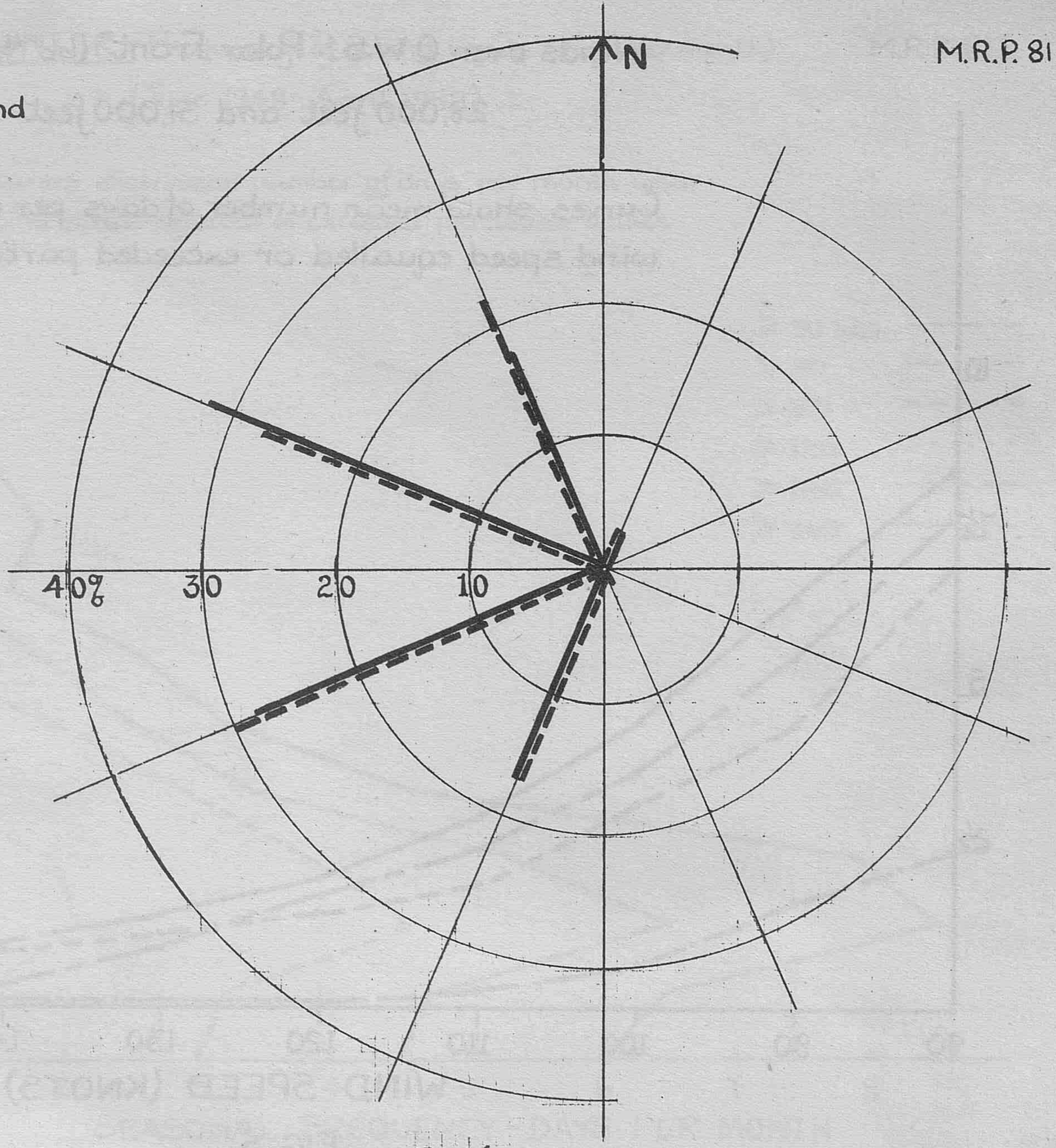


FIG. 6

M.R.P. 812

## SUMMER

Percentage distribution of wind direction when wind was

(a) over 80 knots ——

(b) over 100 knots – – –

45° sectors centred at mid-values

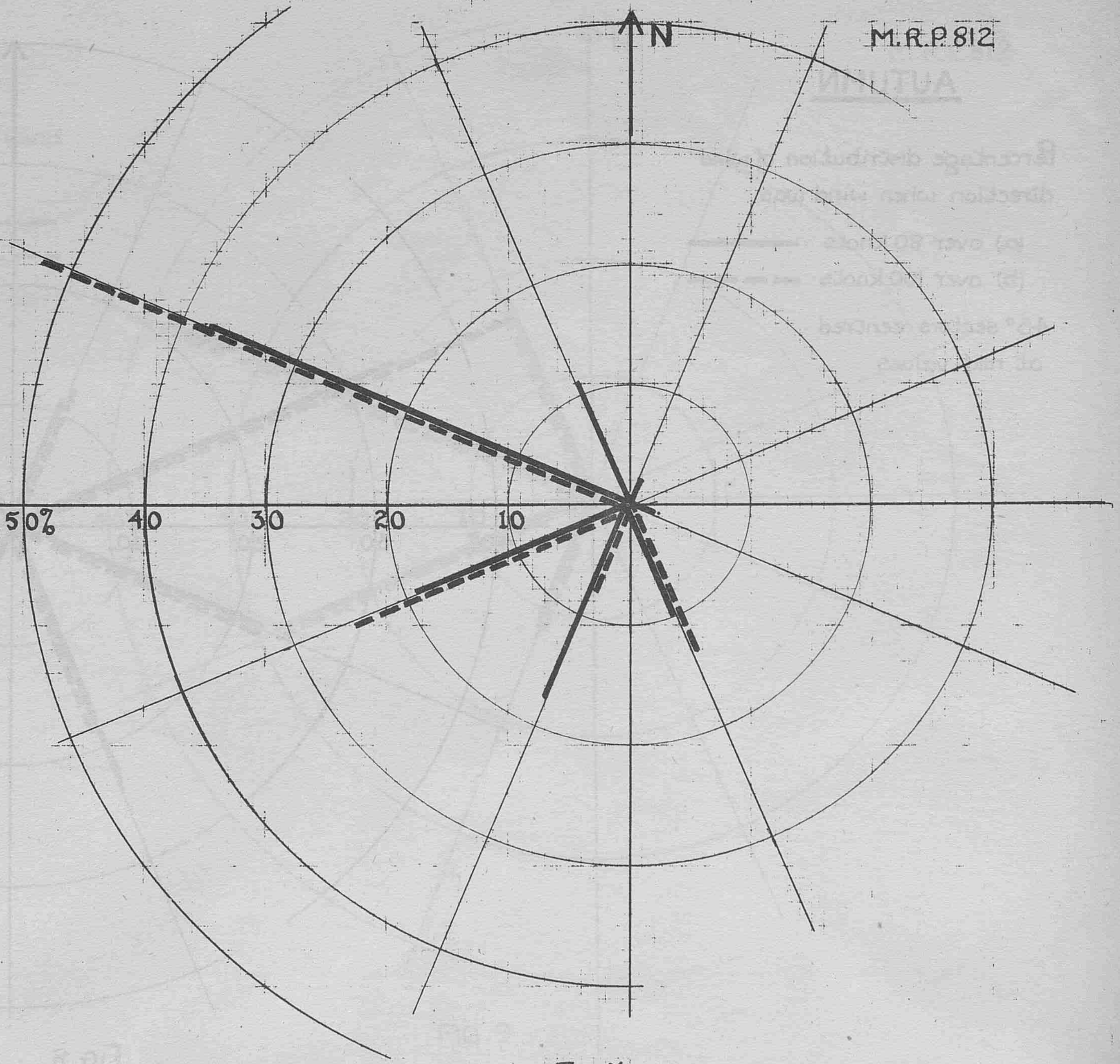


FIG. 7

## AUTUMN

Percentage distribution of wind direction when wind was

(a) over 80 knots ▬▬▬

(b) over 100 knots ▬ ▬ ▬

45° sectors centred at mid-values

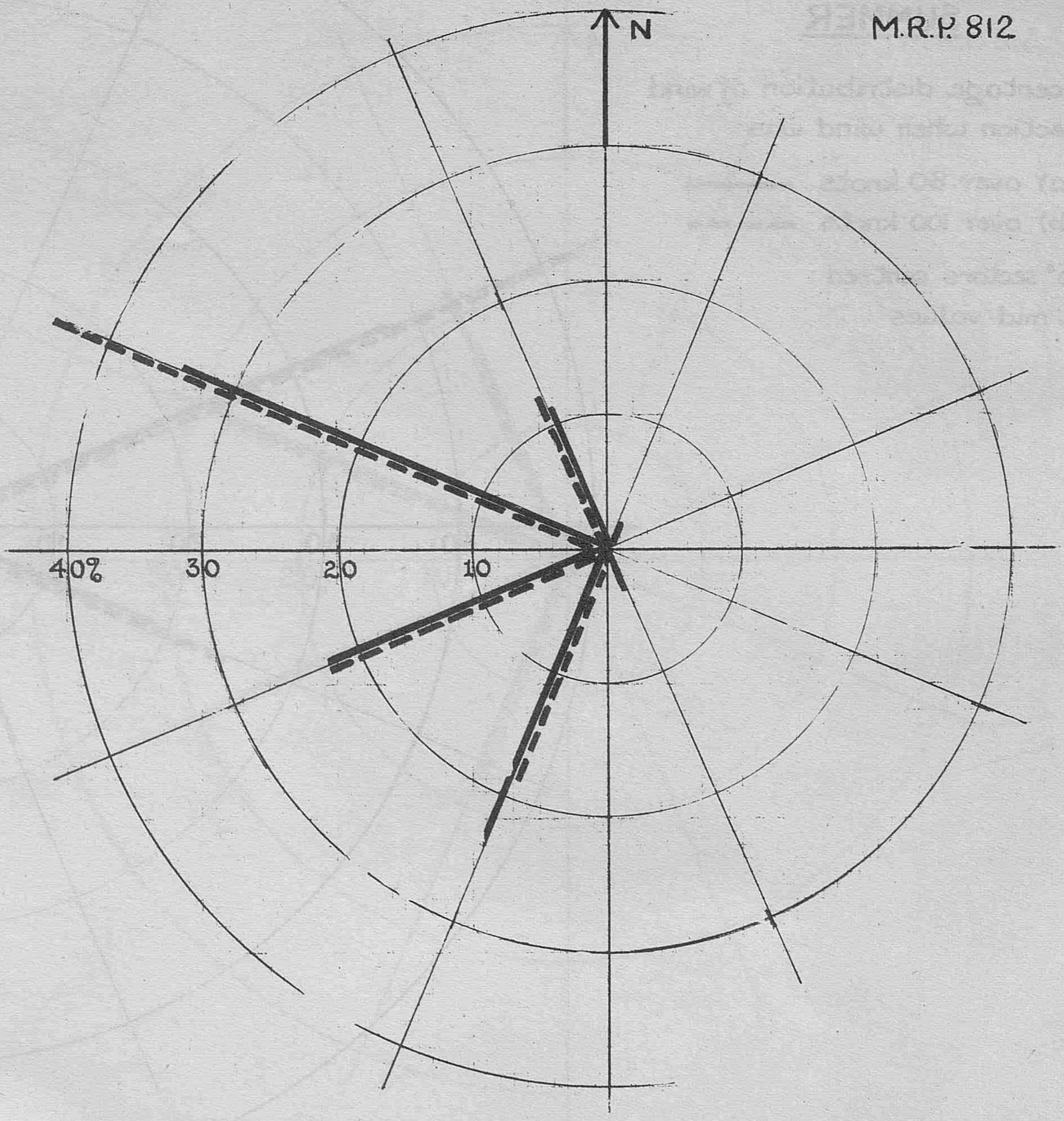


FIG. 8

## WINTER

Percentage distribution of wind directions when wind was

(a) over 80 knots ———

(b) over 100 knots – – – –

45° sectors centred at mid-values

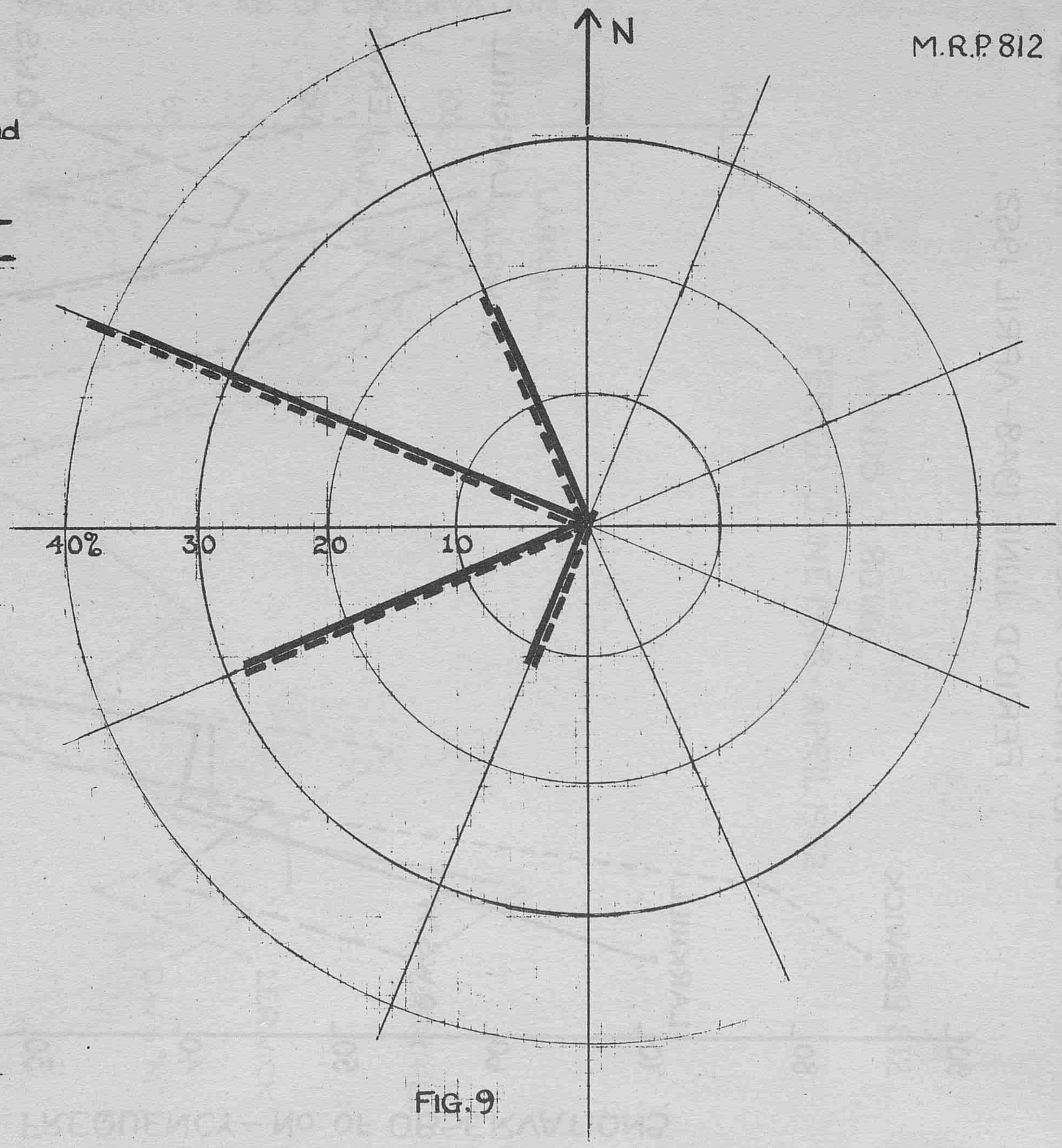


FIG. 9

400 MB. WINDS > 80 kts.

PERIOD JUNE 1948 - APRIL 1952

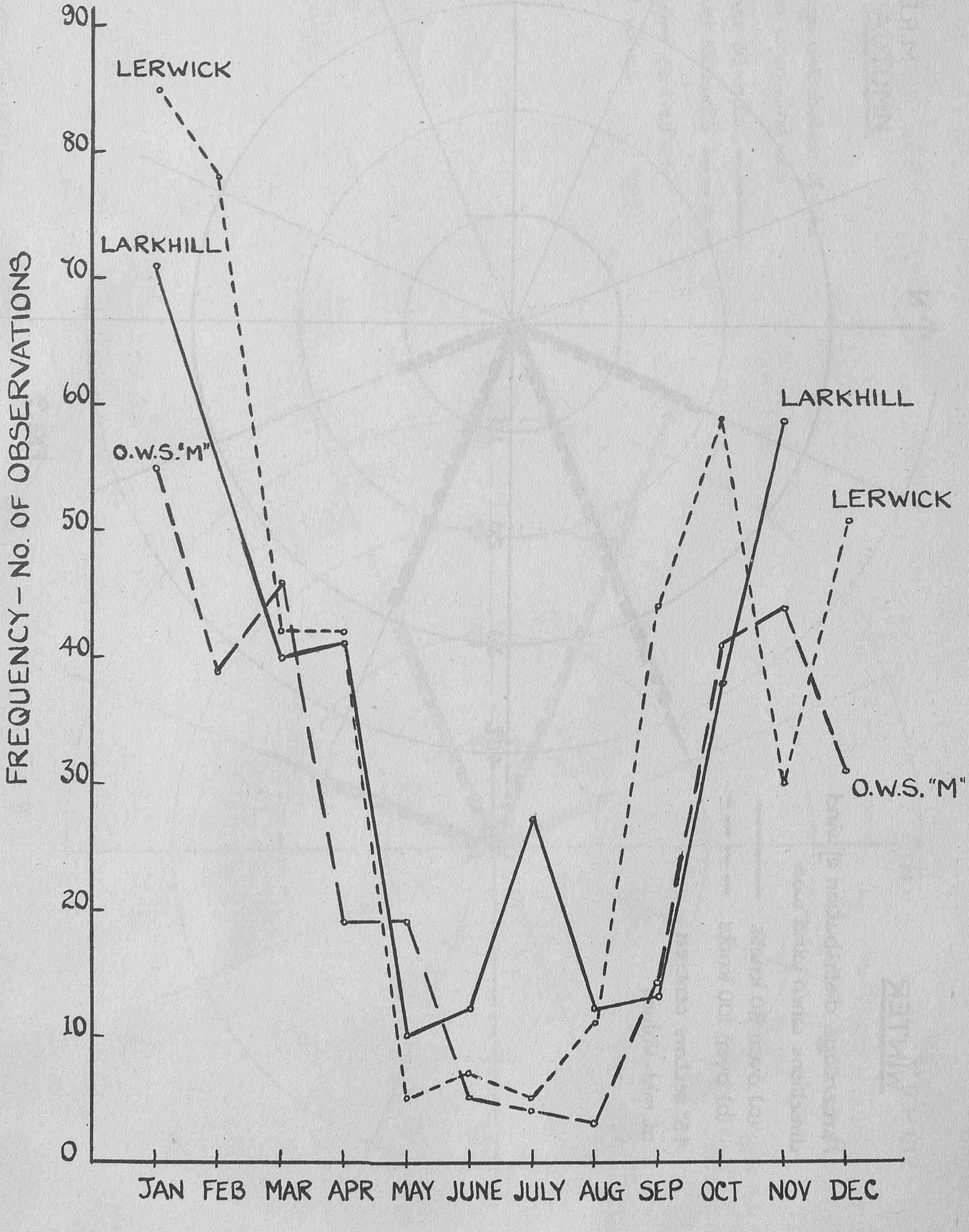


FIG. 10

300 MB. WINDS > 80kts.

PERIOD JUNE 1948 - APRIL 1952

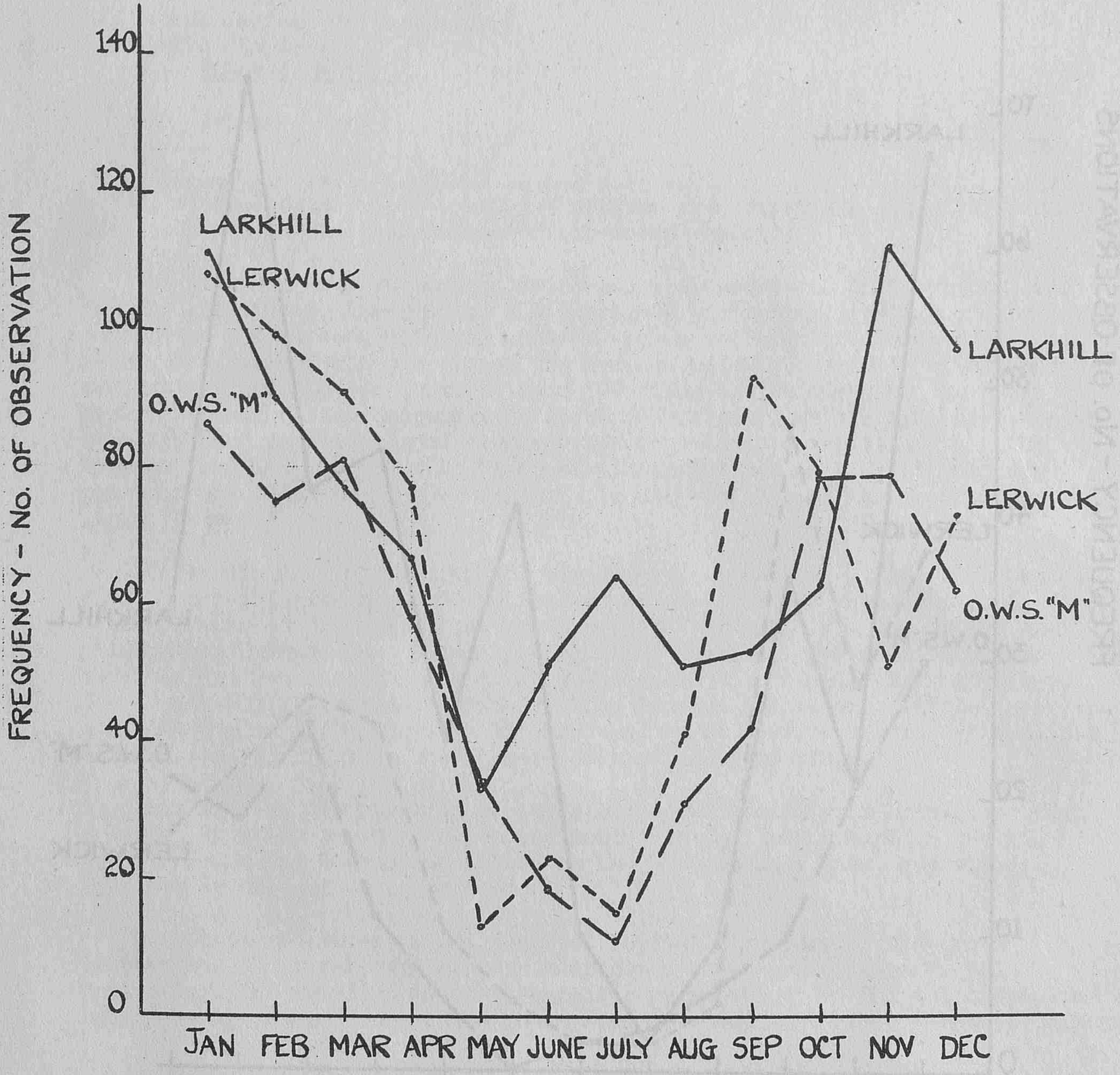


FIG.11

M.R.P. 812

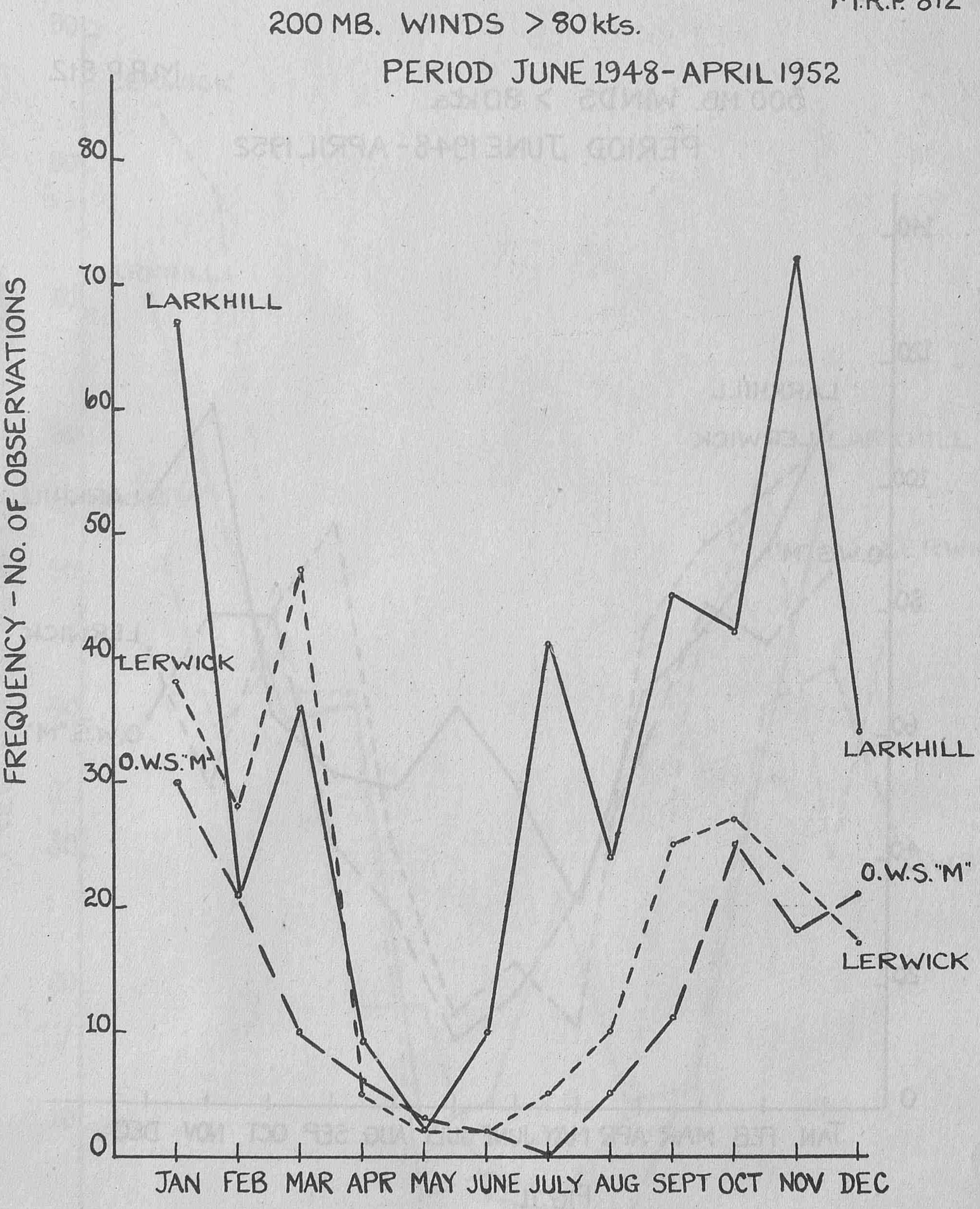


FIG. 12

M. O. 21/120/M. O. 15

Will you please make a further amendment to your copy of M.R.P. 813 as follows:-

Page 15, second paragraph, line 6, amend "flight 7" to "flight 17".

for P Ford(?)
Secretary,
Meteorological Research Committee.

Meteorological Office (M.O.15),
Air Ministry,
London, W.C.2.

9th October, 1953.

G.376987/SEF/10/53/40

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The results of an analyses of [illegible] point, cloud and turbulence near jet streams are presented; brief mention is also made of contrail and indicated air-speed reports.

It is shown from the unique frost-point observations that between about 500 mb. and 200 mb. there is on the average a horizontal gradient of humidity across the jet stream, with the moister air on the high pressure side. At about the level of the jet stream the average relative humidity with respect to ice is around 50 per cent. at some 300 miles on the right of the axis, looking downwind, and decreases to about 10 per cent. at the same distance on the left, but individual jet streams show considerable variability. The average humidity distribution indicates the existence of a relatively dry patch of air below the jet stream axis in the vicinity of the frontal zone at about 500 mb.

Layer cloud of the medium and high type is a feature of the high pressure side of the jet stream, although amounts are variable from case to case: such cloud is rare at distances greater than about 100 n.m. from the jet stream axis on the low pressure side. Layer cloud does not appear to occur above the level of the jet stream axis. The jet streams with fairly extensive layer cloud are usually associated with surface fronts: those with little layer cloud are almost always associated with very weak or no surface front, and such jet streams tend to occur on the forward side of an upper-ridge.

The average distribution of layer cloud and humidity is consistent with a model, in which there is ascending motion in the troposphere on the high pressure side and descending motion on the low pressure side, operating upstream in the jet entrance region.

The broad features of the accepted thermal structure of the upper troposphere and lower stratosphere near jet streams are confirmed. In particular, the temperature observations support the view that the tropopause on the low pressure side of the jet stream is sometimes, but not always, disrupted.

The occurrence of clear air turbulence near jet streams agrees with the results put forward by Bannon (6).

## 1. Introduction

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M.R.P. 812

200 MB. WINDS > 80 kts.

PERIOD JUNE 1948 - APRIL 1952

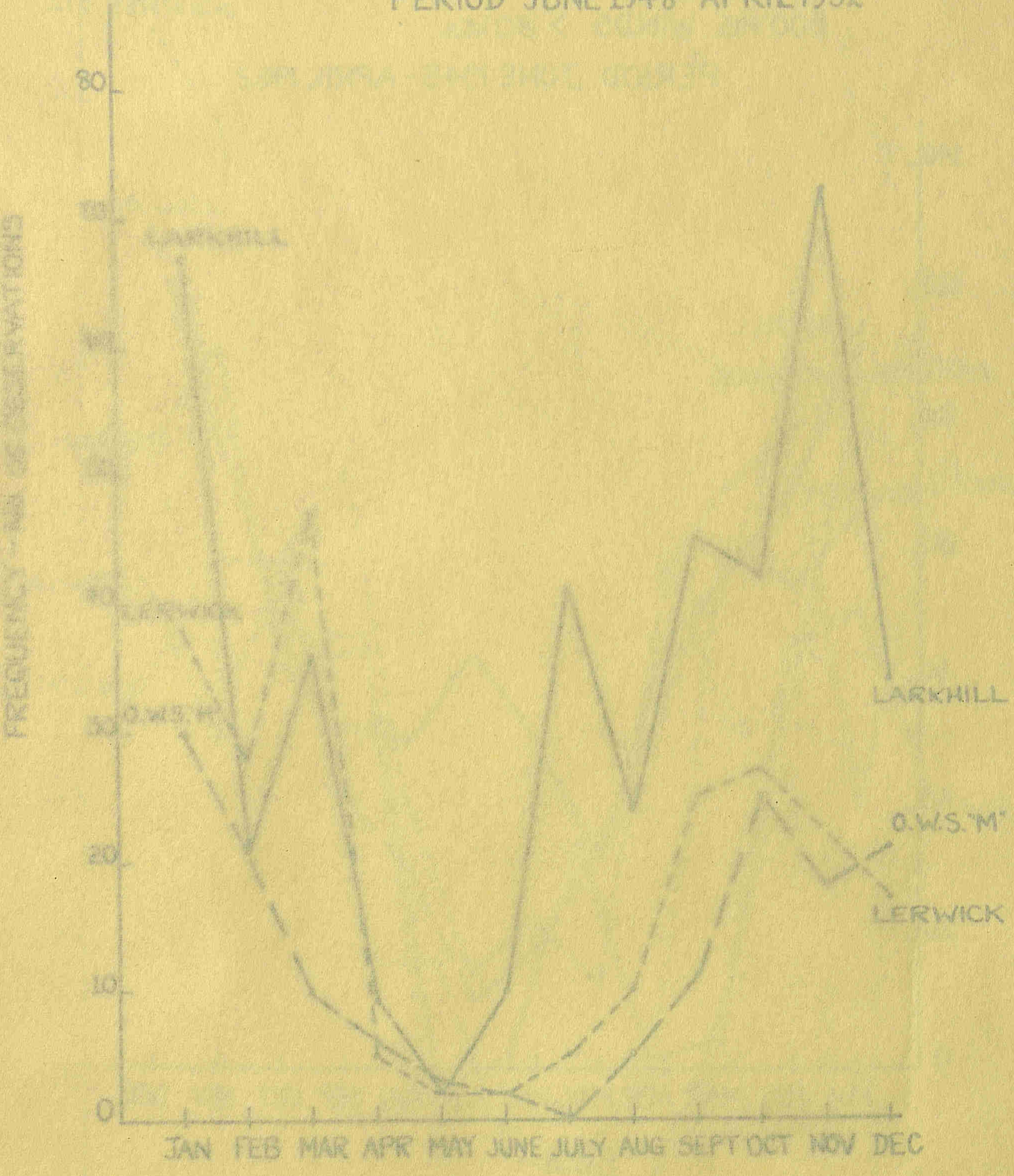

FIG. 12

To members of M.R.C. and of S.C.II

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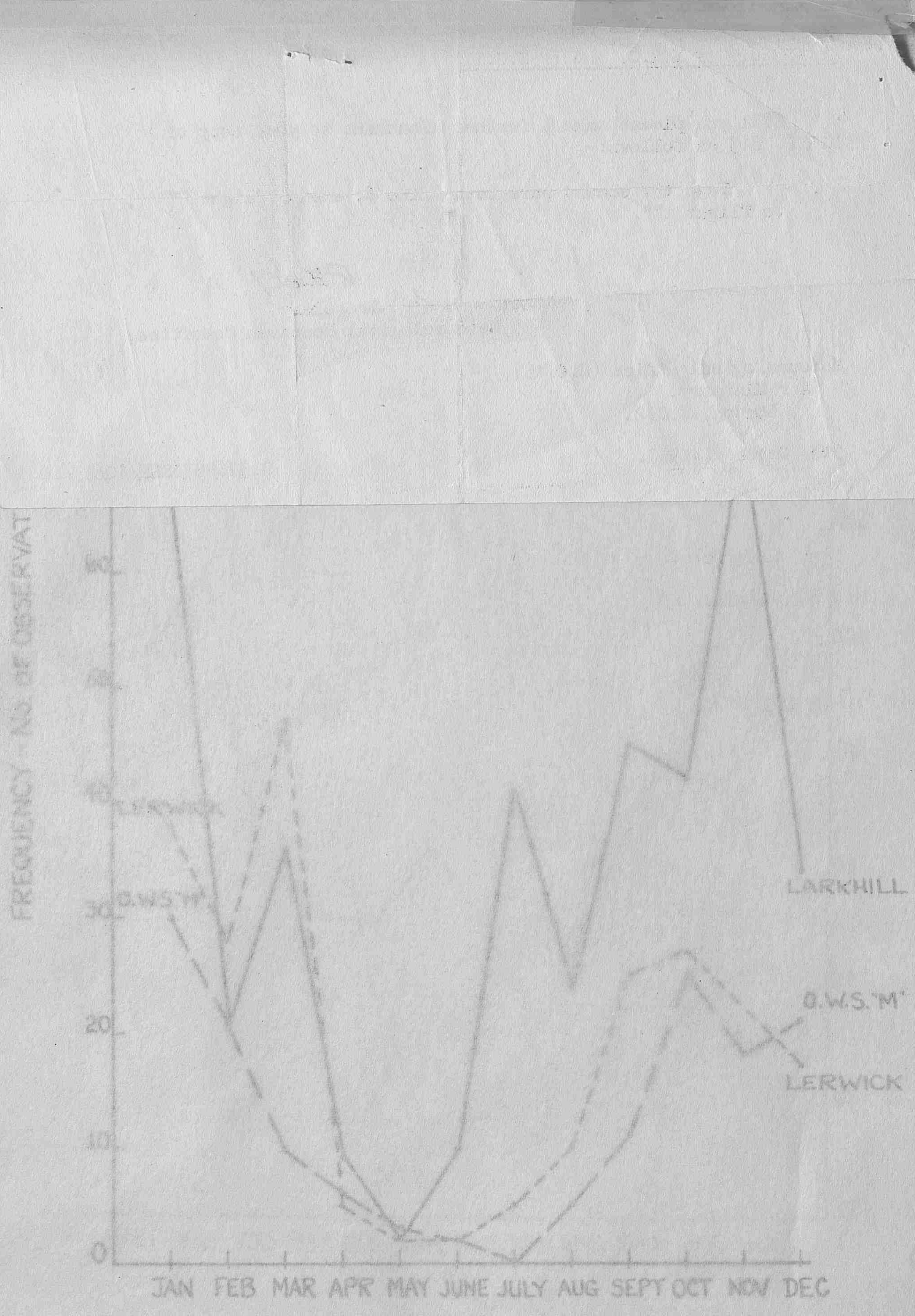
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Fig. 12

"At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 813 by Mr. R. Murray A copy is available in the Library of the Meteorological Office."

M.R.P.813
S.C.II/150
21st May, 1953.

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# The upper troposphere and lower stratosphere near jet streams : an examination of observations made by the Meteorological Research Flight, Farnborough

by

R. Murray (Meteorological Office)

## SUMMARY

The results of an analysis of aircraft observations of temperature, frost-point, cloud and turbulence near jet streams are presented; brief mention is also made of contrail and indicated air-speed reports.

It is shown from the unique frost-point observations that between about 500 mb. and 200 mb. there is on the average a horizontal gradient of humidity across the jet stream, with the moister air on the high pressure side. At about the level of the jet stream the average relative humidity with respect to ice is around 50 per cent. at some 300 miles on the right of the axis, looking downwind, and decreases to about 10 per cent. at the same distance on the left, but individual jet streams show considerable variability. The average humidity distribution indicates the existence of a relatively dry patch of air below the jet stream axis in the vicinity of the frontal zone at about 500 mb.

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## 1. Introduction

Twenty special flights were carried out by the Meteorological Research Flight from May 1951 to August 1952 with a view to investigating the atmospheric structure near jet streams. The central purpose of this paper is to present an analysis of the observations made on these flights. In addition a brief examination is made of some other data of the Meteorological Research Flight; namely (1) the vertical ascents made during the three years 1949-51 within about 400 miles of jet streams, and (2) the observations from the horizontal grid flown in the stratosphere on August 19, 1948. However, unless otherwise mentioned, the statistics and discussions in this paper are generally concerned with the twenty flights made in 1951 and 1952.

On each of these twenty flights vertical grids were flown: the tracks of the horizontal legs (top leg being in reciprocal direction to bottom leg) were chosen to be approximately normal to the jet stream axis. On the horizontal legs various instrumental and non-instrumental observations were made generally every minute, but every half-minute (at least for certain types of observation) on some of the earlier flights. Vertical climb and descent at the end points of the horizontal legs completed the grid: on the vertical legs observations were usually taken every 1000 or 2000 feet. During the climb of the aircraft from Farnborough to the level of the bottom leg, and also later on return to base, some observations were made at rather irregular height intervals; regular and frequent readings were not justified since the extra time consumed in levelling out the aircraft would have meant a shortening of the horizontal legs owing to the limited flight endurance of the aircraft (a Mosquito) and of the meteorological observer. The flights were made in the afternoon or occasionally in the late forenoon, and were usually of duration two or three hours.

The jet streams were selected at the Central Forecasting Office, but in each case the flight plan was agreed upon in consultation with the Senior Meteorological Officer of the Meteorological Research Flight. No rigid criteria were laid down as to what types, or what specific regions or features of jet streams were to be investigated. However, the aircraft was not able to operate at a greater distance than some 200 nautical miles from Farnborough, so that it was necessary to select only those jet streams that were within a few hundred miles of Southeast England. During the period of the experiment the number of cases investigated was limited mainly by the availability of the aircraft; only some weak jet streams were deliberately not investigated after a few of this type had been examined. The specific regions explored were to a great extent determined by the position and orientation of the jet stream axis relative to Farnborough, although there was freedom of choice as regards the levels of the horizontal legs. The length of each horizontal leg of the rectangular grid was usually 150 to 200 n.m.; the bottom leg was at 400 or 350 mb. and the top leg between 300 and 200 mb. in most cases.

These investigational flights have produced an extremely close time and distance network of readings of various meteorological quantities. However, it is emphasised that only limited regions of jet streams were explored on each flight. This has made some of the data difficult to interpret either statistically or specifically in relation to individual jet streams. Moreover, the observations did not include wind, so that wind variations, wind shears and other related phenomena cannot be discussed except by means of the normal synoptic wind reports. Nevertheless, the flight data have been very useful not only in enabling a broad picture of the distribution of humidity and cloud to be obtained, but also in confirming certain other notions about the structure of the free atmosphere near jet streams. If clear-cut, simple models of structure or behaviour of jet streams have not emanated from this study, it is less the result of any inadequacy of data or of the sample of jet streams than the consequence of variety and complexity in nature.

## 2. The flight observations

The one observer on each flight was responsible for making observations of Temperature, Frost-Point, Airspeed, Cloud, Turbulence and Contrails. The methods of obtaining the observations, as well as their frequency and quality, are briefly commented upon below. It should be emphasised that it was a difficult operation for the one observer to maintain a continuous record of all the items listed.

### (a) Temperature

Temperatures, which were reported to the nearest 1.0°F. on most flights but to 0.1°F. on some occasions, were obtained by means of a flat-plate electrical thermometer with an accuracy of around ±0.5°F. under the conditions of these flights. Readings were taken on the horizontal legs generally every minute but every half-minute on some of the early flights, on the vertical legs every 1000 or 2000 feet.

/(b)

(b) Frost-point

Frost-points were measured to the nearest 1.0°F. by means of a Dobson-Brewer frost-point hygrometer; the probable error of these observations is about ±3.0°F. The aim was generally to observe frost-point every minute approximately. However, there were considerable breaks in the readings, partly caused by instrumental trouble and partly due to the difficulty experienced by the observer in maintaining the frost-point records together with the other types of observation.

(c) Airspeed

The indicated airspeed of the aircraft was usually recorded every minute on the horizontal legs. It is estimated that the probable error of the readings is 2 or 3 kt.

(d) Cloud

Visual observations of cloud below and above the aircraft were made at irregular intervals.

(e) Turbulence

The turbulence or "bumpiness" experienced on the flights was estimated and reported, normally every minute, on the scale, 0, 1, 1-2, 2, 2-3, 3 etc.; one unit is roughly equivalent to 0.1 g, where g is the acceleration of gravity. The experienced observers of the Meteorological Research Flight consider that the probable error of these reports is about half a unit or 0.05 g.

(f) Contrails

At convenient intervals notes were made of the occurrence and type of contrails on most of the flights. In straight flight it was not generally possible for the observer to see whether or not contrails were formed.

## 3. Synoptic aspects

At least one vertical cross-section along the flight path - i.e. approximately perpendicular to the jet stream axis - was constructed from the routine aerological observations on each occasion. The cross-sections enabled the flight data to be related to various synoptic features.

Temperatures recorded by the aircraft were generally in fair agreement with values estimated from the cross-section. Perfect agreement cannot be expected since the observations on each flight were necessarily taken at different times, nor were they exactly synchronous with the routine aerological reports. A comparison was made between mean temperatures over 50 n.m. distances on the horizontal legs and interpolated values from the relevant cross-sections to see whether any systematic difference existed. The average temperature difference (aircraft minus radiosonde) was -0.2°F. in the troposphere (above 20,000 feet), 1.1°F. at the tropopause and 1.0°F. in the stratosphere. No allowance for these systematic instrumental differences was made in the analysis in this paper.

The dates of the flights, together with some other facts concerned with the individual jet streams, are listed in the Appendix. The positions of the horizontal legs of the various flights relative to the jet stream axis are conveniently summarised in Fig. 1. The numbers against each leg indicate the relevant flight; only one leg is shown for flight 15, since the aircraft developed engine trouble and the flight was cut short on that occasion; on a few of the flights the bottom leg was actually flown in two parts. No part of any leg was at a greater distance than 250 n.m. on the low pressure side or 450 n.m. on the high pressure side of the jet stream axis. Also, the horizontal legs were all flown within the region from about 14000 feet below to 9000 feet above the level of the jet axis.

The jet streams showed much diversity of type. The wind direction within the core of the jet stream in the region of the cross-section may be classified as follows - 9 cases at or between NW. and N., 7 S. to SW., 2 W. and 2 NE. The wind speed at the jet stream axis averaged 113 kt.; the lowest value was 70 kt. and the highest 150 kt. Some of the jet streams were rather small in longitudinal scale, others were on the scale of the long-waves. The great majority of the jet streams were slow-moving, with only small changes in intensity during the period from 12 hr. before to 12 hr. after the time of the flight: notable exceptions were jet stream 9 which moved quickly eastwards on December 3, 1951, and jet stream 5 which weakened rapidly on June 15, 1951 (the latter case was discussed by Murray (1)). The aircraft explored the central part of the jet stream in most cases; the entrance and exit regions were not investigated except perhaps on the occasion of flight 6 (near marked confluence in wind flow).

## 4. Some thermal features

### (a) Temperature irregularities

The temperature readings on the horizontal legs of the flights made by the M.R.F. aircraft invariably showed small amplitude fluctuations over short time intervals. The magnitude of these effects may be measured by the standard deviation of the individual observations from a 9-minute running mean (over about 45 n.m.), or by the r.m.s. temperature variation over one-minute intervals. These two quantities were computed for the individual legs which were classified into the three categories (1) leg entirely in the troposphere, (2) leg partly in troposphere and partly in stratosphere and (3) leg entirely in stratosphere. All the observations pertaining to each of these three categories were combined with the result shown in Table I.

Table I Standard deviation of individual temperature readings from 9-minute running mean and r.m.s. one-minute temperature variation for observations in upper troposphere, near tropopause and in lower stratosphere.

| Type of temperature run | Standard deviation (°F) | r.m.s. one-minute variation (°F) |
|---|---|---|
| entirely in troposphere (at or above 400 mb.) | 0.55 | 0.70 |
| partly in troposphere and stratosphere | 0.81 | 0.98 |
| entirely in stratosphere | 1.05 | 1.23 |

Table I shows quite clearly that the temperature irregularities are significantly greater in the stratosphere than in the upper troposphere. The same general trend is displayed in Fig. 2 which shows the standard deviations for the individual legs plotted against the pressure-level of the leg. Examples of the temperature irregularities observed in the troposphere and in the stratosphere are shown in Fig. 3, which refers to the flight on February 1, 1952: Fig. 3 contains relevant surface, 300 mb. and cross-section charts which enable the general synoptic picture to be readily seen.

However, Frith (2) has already pointed out that temperature fluctuations of this type normally decrease in intensity with height in the troposphere and increase again on entering the stratosphere: the present series of observations gives an interesting confirmation of this fact which is not specially a feature of jet streams.

### (b) Temperature profiles near jet streams

The temperatures observed on the horizontal legs do not add much to existing ideas concerning the thermal structure of the upper troposphere and lower stratosphere in the vicinity of jet streams. The picture presented by the cross-sections constructed from routine upper-air data was generally confirmed except in points of detail.

On each leg the temperature difference between the end points was noted and the mean temperature gradient (in terms of °F. per 100 n.m.) thereby computed. From the relevant sections estimated temperature gradients were also made. The mean absolute value of the temperature gradient based on all the legs was 3.0°F. per 100 n.m. compared with a mean value of 3.5°F. per 100 n.m. obtained from the sections. The mean value, irrespective of sign, of the difference of the magnitudes of the temperature gradients obtained by the two methods was 1.4°F. per 100 n.m. Furthermore, the direction of the overall temperature gradient invariably agreed with that interpolated from the section. Thus the broad scale picture given by the cross-section was generally confirmed.

The broad tropospheric baroclinic zone, generally associated with jet streams and extending horizontally over several hundred miles, was not completely traversed on any one flight. Some part of 12 legs (i.e. bottom legs of flights 1, 3, 6, 8, 10, 12, 13, 14, 17, 18 and 19 and top leg of flight 8 - see Fig. 1) was located vertically below the jet stream axis. The readings obtained on these legs may be regarded as representative of conditions within the baroclinic zone in the upper part of the troposphere. The individual temperature gradients on these 12 legs varied from 1.0°F. per 100 n.m. to 7.0°F. per 100 n.m. with a mean value of 3.7°F. per 100 n.m. Not unexpectedly, the smaller temperature gradients tended to be associated with the weaker jet streams and with the flight legs nearer the level of the jet stream axis. The temperature profile generally showed a fairly smooth trend with superimposed temperature fluctuations of the size mentioned in section 4(a). A typical example is shown in Fig. 4 by the temperature profile observed at 400 mb. on April 24, 1952. In passing, it will be noted that the temperature gradient on the leg at 250 mb. (above the level of the jet stream axis but in the troposphere) was in the opposite sense to that at 400 mb.: such a reversal of the sign of the general tropospheric thermal gradient in the few thousand feet which often separates the jet stream axis from the tropopause is frequently suggested on cross-section charts, and is of course a necessity if the upper flow is quasi-geostrophic.

Clearly defined frontal zones in the troposphere at or above 400 mb. were not generally indicated by the flight observations although it should be remembered that most of the legs were not located in the region where fronts might be expected. The only well defined front was the cold front associated with the SSW. jet stream of June 18, 1951 (flight 6), of which the synoptic situation and temperature profile are shown in Fig. 5.

/The

The temperature profile at 400 mb. clearly shows the frontal boundary adjacent to the warm air within which the horizontal temperature gradient is smaller than suggested on the cross-section.

It might be expected that an aircraft investigation near the level of the jet stream axis and on the low pressure side would throw some light on the question of the continuity of the tropopause in that region. Flight 2 on May 17, 1951, was made with that end in view. The relevant synoptic charts and cross-section are shown in Fig. 6. The cross-section through the NE. jet stream suggested a definite break in the tropopause although the evidence was inconclusive. The temperature observations taken by the aircraft (see Fig. 6) confirmed that at this time there was no indication of a tropopause in the region examined. No other flight was quite so satisfactorily situated for the specific purpose of examining the nature of the tropopause just to the low pressure side of the jet stream axis, although several traversed the tropopause. In some cases the tropopause was drawn on the cross-section as continuous from the high to the low pressure side of the jet stream, in others it was drawn as uncertain or discontinuous, but in none of these cases did the flight observations add any additional evidence to permit a definite or a contrary construction to be made. It might be mentioned, however, that the tropopause tended to be continuous with the weaker jet streams, and to be uncertain or discontinuous with the more intense ones.

### (c) Temperature pattern on horizontal grid flown on August 19, 1948

It is desirable to put on record the data obtained on a flight made between 1000 and 1200 G.M.T. on August 19, 1948. Readings were taken on level flight at 35,000 feet over a roughly rectangular area relative to the air. This area was traversed six times along nearly parallel legs running south to north, each some 60 mi. long and about 6 mi. apart. The first leg commenced at Farnborough and finished 60 mi. farther north, and each leg was 6 mi. west of the previous one relative to the air. On this occasion the air flow in the upper troposphere and lower stratosphere was WNW'ly. The position of the jet stream axis cannot be determined with any precision owing to lack of wind observations over France, but the observations over England indicate that the wind maximum was to the south of the area over which the flight was made. The flight data refer to an area relative to the air at 35,000 feet in the lower stratosphere on the low pressure side of the rather weak WNW. jet stream. The synoptic situation is illustrated in Fig. 7 together with the temperature and frost-point patterns. Fig. 7(d) is of course a plot of non-synchronous temperature readings relative to the air; because of the horizontal wind shear the plot should not be in the form of the rectangular grid shown, but it is difficult to construct the true air-grid.

The temperature readings on the 60 mi. legs were made every 3 mi. or so and show irregularities of the type mentioned in section 4(a). However, perhaps the main point of interest is the existence of quite a pronounced horizontal temperature gradient in a direction from NE. to SW. in the stratosphere above and slightly on the low pressure side of the jet stream axis. The true direction of the temperature gradient is probably from NNE. to SSW. but is distorted on the diagrammatic representation because of horizontal wind shear.

/5.

## 5. Humidity distribution near jet streams

### (a) Frost-point observations on the 20 special flights

Frost-points were more variable than temperatures. Over intervals of 2 to 5 min. (horizontal distances 10 to 25 n.m.) fluctuations greater than 10°F. occurred on 11 legs, and greater than 20°F. on 2 legs. On 11 legs with fairly complete sets of readings at one minute intervals the r.m.s. one minute variations of frost-point ranged from 1.0°F. to 3.3°F.; these fluctuations tended to be rather greater for the legs entirely in the troposphere. Frost-point fluctuations of this type have already been studied by Frith (3) and are not particularly a feature of the atmosphere near jet streams.

Vertical frost-point sections were constructed in most cases for the grid region flown by the aircraft. The observations were not adequate to show all the minor variations of frost-point within the grid region but the broad picture was usually delineated. Two such frost-point sections are presented in Fig. 8 (flight 3 on May 24, 1951) and Fig. 9 (flight 12 on January 2, 1952). On 11 occasions (flights 2, 3, 4, 6, 10, 11, 12, 13, 14, 17 and 19) it was found possible, with the aid of these frost-point sections, to estimate the approximate values of temperature, frost-point and frost-point depression (i.e. temperature minus frost-point) at the jet stream axis. The mean values were -60°F. for temperature, -79°F. for frost-point and 19°F. for frost-point depression. The frost-point depression varied from about 0°F. (flight 12) to 43°F. (flight 14); although these figures are uncertain to within 5°F. at least, they do indicate the existence of great differences in humidity at the jet stream axis from case to case; indeed it appears that the air in the core of the jet stream can be practically saturated or quite dry. However, it is noteworthy that there is an association between a high value of frost-point depression (i.e. low relative humidity) and high temperature in the core of the current. For example, the five jet streams (cases 2, 6, 14, 17 and 19) with relatively warm air at the position of the jet axis had a mean temperature of -50°F. and a mean frost-point depression of 32°F.; in contradistinction, the other six cases (flights 3, 4, 10, 11, 12, 13) averaged -68°F. for temperature and 9°F. for frost-point depression in the core of the current. The jet streams with a relatively moist core and those with a relatively dry core do not appear to have any particular seasonal preferences, nor to be related to broad synoptic features such as the forward side of a long-wave trough or ridge, nor to be connected with the direction of lateral motion of the axis; but the smallness of the sample makes it uncertain whether these results are truly representative of jet streams.

The frost-point sections in Figs. 8 and 9 indicate a horizontal gradient of frost-point in the same direction as the pressure gradient. This type of frost-point distribution was suggested on many of the flights, despite the masking of the effect on some occasions by irregularities from the general trend or by missing observations. An example of a different frost-point distribution is shown in Fig. 6: the overall horizontal gradient of frost-point along the leg at 275 mb. is practically zero, and along the bottom leg at 325 mb. it is in the opposite sense to the horizontal pressure gradient. However, the facts that the region explored by the aircraft on individual occasions was always limited in extent and differed in position relative to the jet stream axis, as well as certain shortcomings of the frost-point observations, made it necessary to combine the data from the different flights in order to bring out the overall features of the frost-point distribution. Various combinations of the data are discussed below.

In the first place it seems quite clear that there is normally a horizontal gradient of frost-point from the troposphere to the stratosphere. Seven legs which quite certainly traversed the tropopause, namely those at 250 mb. on flights 3, 4, 7, 12, 19 and 20 and at 220 mb. on flight 19, were employed in the construction of Fig. 10 which shows the mean horizontal variation of temperature and frost-point relative to the tropopause values: the considerable difference in humidity between the stratosphere and the troposphere at the same level is obvious.

There were 18 legs (flights 3, 5, 6, 8, 10, 12, 13, 14, 17, 19) with fairly complete frost-point readings and such that the frost-point observation was available or could be estimated with confidence on each leg at the point vertically above or below the axis of the jet stream. The frost-point was then obtained on each leg at positions 50 n.m. apart on each side of the jet stream axis, and hence the frost-point relative to that at the point of intersection of the leg with the vertical through the jet stream axis. The averages of these relative frost-points at different distances for the legs above the level of the jet stream axis are displayed in Fig. 11(a), and those for the legs below the jet stream axis in Fig. 11(b). The frost-point profile in Fig. 11(a) indicates a horizontal gradient of frost-point across the jet stream in the same direction as the pressure gradient. Six of the seven legs employed in obtaining Fig. 11(a) were either in the uppermost part of the troposphere or partly in the troposphere and partly in the stratosphere, and these six legs individually agreed with the mean profile: the remaining leg, that of flight 13, was entirely in the stratosphere and had frost-points at 100 n.m. and 150 n.m. on the high pressure side some 10°F. lower than the values at the points 0 and 50 n.m. The average frost-point profile for the legs below the jet stream axis, shown in Fig. 11(b), is of the same type as that shown in Fig. 11(a), apart from the zero value at 100 n.m. on the high pressure side. Three of the eleven legs used in constructing Fig. 11(b) had frost-point gradients agreeing with the mean picture: the other eight legs, taking a broad view, showed a similar overall trend but with frost-point irregularities so large as to give a contrary trend over substantial parts of each leg. Evidently there is a definite tendency for the air to be moister, as measured by higher frost-points, in the region 100-200 n.m. to the right compared with the region 100-200 n.m. to the left of the jet stream axis, looking downwind; this type of humidity distribution appears to occur both above and below the level of the jet stream axis. The profiles of temperature and frost-point shown in Fig. 10 indicate that a substantial humidity gradient exists from the troposphere to the stratosphere at the same level, and this may account for much of the transverse humidity gradient shown in Fig. 11(a) for the legs above the axis of the jet stream.

In order to construct some kind of humidity profile across a larger extent of the jet stream, the frost-point depressions (i.e. temperature minus frost-point) were obtained from all the legs which were flown entirely in the troposphere. Fig. 12(a) shows the mean picture of the depression of the frost-point across the jet stream, based on 23 legs in the upper troposphere. The jet streams were subdivided into two groups, viz. (1) those which either were quasi-stationary on the forward side of a cold upper-trough or moved normal to the axis in the direction from cold to warm troposphere and (2) those which either were quasi-stationary on the forward side of a warm upper-ridge or moved normal to the axis in the direction from warm to cold troposphere. The mean profile of the frost-point depression obtained from the group (1) cases is shown in Fig. 12(b), that from group (2) cases in Fig. 12(c). The profile of Fig. 12(b) is unfortunately rather limited in extent, but it does not appear to be radically different from the profile of Fig. 12(c) over a similar distance relative to the jet stream axis. There is thus little or no evidence to suggest that the profile shown in Fig. 12(a), that is with a definite horizontal gradient of humidity in the upper troposphere across the jet stream from high to low pressure, is, or is not, greatly dependent upon the broad synoptic type of jet stream.

Confirmation of the tendency for a horizontal gradient of humidity across jet streams will be given in the following section by an analysis of independent data. The tropospheric humidity distribution will be shown in section 6 to fit in with the average distribution of layer cloud near jet streams. Also in section 6 the significance of the humidity and cloud distribution in relation to vertical motion will be discussed.

### (b) Mean humidity distribution obtained from vertical soundings made in 1949-51

Those vertical soundings made by the Meteorological Research Flight during 1949-51 within 400 miles of the axes of jet streams were examined. For this purpose a jet stream was taken as a wind maximum of at least 70 kt. on the 300 mb. charts constructed by the Central Forecasting Office. The soundings were categorised according to their position relative to the jet stream axis (as estimated on the 300 mb. chart): the four groups comprised observations within 200 mi. and within 200-400 mi. from, and on either side of, the jet stream axis. Mean values of temperature, frost-point and relative humidity with respect to ice were computed in each category for pressure levels at 50 mb. intervals from 500 mb. to 200 mb. Some soundings either stopped between 500 mb. and 200 mb. or had an incomplete set of frost-point readings. These were discarded from the analysis if less than five humidity observations (at 50 mb. intervals) were available: eight soundings were thus rejected: but in only one of these cases were the frost-point observations missing because the aircraft was definitely in cloud (cirrostratus), and this flight took place some 300 mi. from the jet stream axis on the high pressure side. Altogether 56 vertical soundings were available for the analysis: on the low pressure side there were 6 at 200-400 mi. and 16 at 0-200 mi., on the high pressure side 14 at 0-200 mi. and 20 at 200-400 mi. from the jet stream axis.

In Fig. 13, diagram (a) is a composite cross-section, based on the 56 soundings, which shows the temperature and frost-point distribution; diagram (b) shows relative humidity with respect to ice in place of frost-point. The humidity pattern above 400 mb. could be a reflection of the preponderance of stratospheric observations on the low, and of tropospheric observations on the high, pressure side of the jet stream axis; for the air is drier in the stratosphere than in the troposphere at the same level, as has been illustrated in Fig. 10. The overlap between stratospheric and tropospheric observations occurred between 400 and 300 mb. in the group at 200-400 mi. from the axis on the low pressure side, and between 250 and 200 mb.

at 200-400 mi. from the axis on the high pressure side. Another composite diagram - Fig. 13(c) - was constructed so as to emphasise the difference in humidity between stratospheric and tropospheric air. The top left of Fig. 13(c) is based on mean values from individual soundings entirely in the stratosphere (marked S); the remainder of this figure refers to tropospheric conditions. The criterion for deciding that the points S were in the stratosphere was simply the existence of a majority of stratospheric observations at these points - the tropospheric observations at the same points being then neglected in evaluating the mean temperature and humidity: a similar criterion applied to the tropospheric points. Fig. 13(c) shows that the main part of the horizontal gradient of humidity which is indicated in the upper parts of Fig. 13(a) and (b) is concentrated near the tropopause.

There are clearly two main features of the average humidity distribution shown in Fig. 13:-

(1) Above about 400 mb. the air tends to be relatively dry on the low pressure side and relatively moist on the high pressure side of the jet stream. The horizontal gradient of humidity is probably, in the main, a result of the fact that, at about the level of the wind maximum, the air on the low pressure side is usually stratospheric, whereas on the high pressure side it is tropospheric. Nevertheless, there also appears to be a tendency for a horizontal humidity gradient to exist in the upper troposphere; a similar feature has been noted in connection with the analysis of the 20 special flights.

(2) At or below about 500 mb. a relatively dry patch tends to occur roughly beneath the jet stream axis. It is clear that this relatively dry patch has a close association with the frontal zone which usually intersects the 500 mb. surface directly beneath the jet stream maximum. Sawyer (4) has confirmed the existence of a dry patch near many frontal zones: he suggests that its presence can be accounted for by the advection into the frontal zone of air which has been dried by subsidence at an earlier stage in a region external to the frontal zone.

### (c) Frost-point pattern on horizontal grid flown on August 19, 1948

The air-plot of the frost-point observations made on a horizontal flight in the stratosphere near a weak jet stream on August 19, 1948 is shown in Fig. 7(e); the corresponding temperature observations have been discussed in section 4(c). Fig. 7(e) is of interest in illustrating the fluctuations of frost-point that occur in the stratosphere, and in showing the existence of a gradient of frost-point in the stratospheric air associated with the jet stream.

## 6. Cloud distribution near jet streams

The observations of cloud from the 20 special flights have enabled some broad features of the cloud distribution near the jet stream to be made fairly clear, although it was not found possible to present in any detail the cloud structure near individual jet streams, except occasionally over quite small regions.

Cloud was actually traversed on the horizontal legs rather infrequently, as is indicated by Table II which gives the percentage of the flying time in cloud of various types. The 39 legs shown in Fig. 1 occupied some 26½ hours flying time.

/Table II

Table II Percentage of flying time on horizontal legs when aircraft was in cloud of various types.

| | Position of legs relative to jet stream axis | | |
|---|---|---|---|
| | Above | Below | Above and below |
| | | percentage | |
| Cirrostratus, cirrus, anvil cirrus | 0 | 2.9 | 1.8 |
| Nimbostratus, altostratus, altocumulus | 0 | 3.6 | 2.2 |
| All cloud | 0 | 6.5 | 4.0 |

Not only was no cloud traversed on any of the 16 legs which were located above the jet stream axis, but only on three of these legs (flights 9, 10 and 16) were cirrus or cirrostratus patches observed above the aircraft. The cirrus type of cloud occurred at distances varying between about 30 n.m. and 400 n.m. from the jet stream axis on the high pressure side in the upper troposphere; no cloud was observed at or above the level of the jet stream axis on the low pressure side. It seems clear that the atmosphere above the jet stream axis is generally cloudless, apart from occasional patches of cirrus or cirrostratus on the high pressure side in the uppermost part of the troposphere.

Seven legs (flights 1, 3, 7, 9, 12, 13 and 16) traversed some cloud. The one report of cumulonimbus anvil occurred just over 200 n.m. from the axis on the low pressure side of jet stream 7, below which region a good deal of convection was taking place. The delicate cirrus which was traversed on flight 13 occurred practically vertically below the jet stream axis. However, on the remaining five flights the aircraft traversed cloud of the layer type, i.e. nimbostratus, altostratus, altocumulus and cirrostratus, which occurred at distances from the jet stream axis varying from about 70 n.m. on the low to 400 n.m. on the high pressure side. As illustrative examples may be cited the flights 3 and 12, the synoptic situations of which are shown in Figs. 8 and 9: in each case layer cloud was encountered in the region corresponding to the bottom right hand corner of diagrams (d), i.e. beneath and to the high pressure side of the axis of the jet stream.

On seven flights (numbers 2, 4, 5, 8, 11, 15 and 17) no cloud was observed at or above the bottom leg. In each case there was usually broken stratocumulus or cumulus below the aircraft, and on one or two occasions some local altocumulus. These seven jet streams differed considerably in both intensity and orientation. However, it is noteworthy that the region investigated extended from about 230 n.m. on the low to about 150 n.m. on the high pressure side of the jet stream axis; in other words the legs were somewhat biased towards the low pressure side of the jet stream axis. Although the position of the region investigated almost certainly was associated with the almost complete absence of medium and high cloud on these particular flights, it is also significant that in six of the cases no surface front accompanied the jet stream; the typical sloping baroclinic zone appeared linked in each of these six cases (all but number 5) with an inversion or very stable layer of the subsidence type in the lowest few thousand feet. The other jet stream at the time of flight 5 was rapidly weakening and was associated with a weak surface cold front of the kata type: the sequence of events associated with this case has already been discussed in M.R.P. 743 (1).

In contrast with the seven flights with no cloud at or above 400-350 mb., the remaining 13 flights reported some cloud at (seven flights) or above (six flights) the bottom leg. These 13 flights ranged from nearly 250 n.m. on the low to just over 400 n.m. on the high pressure side of the jet stream axis.

Excluding the nimbostratus, altostratus and altocumulus cloud actually traversed and already discussed, all the cloud was ice crystal type, i.e. cirrus, cirrostratus or combination with some cirrocumulus in addition on one occasion; excluding also the cumulonimbus anvil cloud traversed at just over 200 n.m. from the axis on the low pressure side of the jet stream, the high cloud was observed from about 70 n.m. on the low to just over 400 n.m. on the high pressure side of the axis of the jet stream, i.e. predominantly on the high pressure side. The jet streams investigated by these 13 flights cannot easily be placed into convenient synoptic pigeon-holes: they were situated on the forward side of both upper-troughs and upper-ridges; they were quasi-stationary or mobile; and they were diverse as regards intensity and synoptic size. However, the most notable synoptic feature associated with these particular jet streams was the existence of a surface front in all but one or two cases (e.g. no surface front on May 3, 1951), although more than half of the fronts were quite weak in terms of rainfall activity.

Collating the observational evidence from the entire series of flights, it appears that the cloud in the atmosphere above about 400 mb. is predominantly on the high pressure side of the jet stream axis. More precisely, above 400 mb., it may be said: first, that generally no cloud (except occasionally cumulonimbus tops or anvils) occurs on the low pressure side within the region between about 250 n.m. and 100 n.m. from the jet stream axis; secondly, that some cloud generally occurs on the high pressure side between about 150 n.m. and 450 n.m. from the jet stream axis; and, thirdly, that from about 100 n.m. on the low to about 150 n.m. on the high pressure side there may be nil or any amount of cloud from one jet stream to another.

So far the discussion has centred on the cloud distribution at or above 400 mb. On most of the flights some cloud was generally observed below the aircraft. As regards low cloud, it was predominantly cumulonimbus or cumulus or stratocumulus on the low pressure side, and stratocumulus or nimbostratus on the high pressure side of the jet stream axis, but amounts were variable from case to case: such a distribution of low cloud was generally supported also by the surface synoptic observations, and is not unexpected. Medium cloud tended to be much more common when the region explored was on the high pressure side of the jet stream axis, as with cloud above 400 mb.

Quite clearly the distribution of layer cloud (excluding low cloud) near jet streams appears to be of most interest and to fit into a rough pattern. Therefore, it was decided to form composite pictures from all the aircraft observations of layer cloud, irrespective of whether the cloud was observed above or below (or at) the level of the flight, in order to bring out the main features of the cloud distribution. For this purpose the jet stream was divided into sectors of width 100 n.m., as shown on the abscissae of Figs. 14 and 15; the fraction of each sector examined on each flight, whether or not layer cloud was observed and, if so, over what fraction of the sector, were all noted, thus enabling the overall distribution of layer cloud to be presented. Fig. 14 shows the percentage frequency of occurrence of layer cloud (whatever the amount) within the various sectors; histogram (a) is based on reports of nimbostratus, altostratus and altocumulus only, whereas (b) includes also cirrostratus type. Fig. 15 presents the observational data rather differently; it shows for each sector the observed cloud amount expressed as a percentage of the maximum amount of cloud that would have been observed if the flights had reported layer cloud, all the time, either above or below the grid track. Despite the smallness of the sample upon which the histograms of Figs. 14 and 15 are based, all tell a consistent story. The placing of cloud observations on individual flights in their correct position relative to the jet stream axis was uncertain in some degree; but there is no doubt of the reality of the overall distribution of layer cloud portrayed in Figs. 14 and 15. It is quite clear that layer cloud is mainly a feature of the high pressure side of jet streams, although great variations in individual cases make one hesitant to say that the average picture is the typical model in any but broad respects.

Examination of the surface synoptic and the aircraft observations suggests that the jet streams with extensive layer cloud mainly on their high pressure side are usually associated with a surface front which may be either active or rather inactive in terms of weather; such jet streams may be located on the forward side of either an upper-trough or an upper-ridge. Moreover, the jet streams with little or well broken layer cloud on their high pressure side are normally either not associated with a front on the surface chart (a low level subsidence inversion usually occurs where a front might be expected) or the front is very weak; these jet streams occur more frequently on the forward side of an upper-ridge than on the forward side of an upper-trough.

The distribution of layer cloud is in agreement with the type of humidity distribution discussed in section 5(a) and (b). The absence of fairly extensive layer cloud is not related in any simple way to general descending motion, although there is probably some positive correlation between the two factors. However, it is quite clear that, in order to produce such layer cloud, there must have been ascending motion of the air, although it may have ceased at the time or at the place of observation of the cloud. Thus the average distribution of layer cloud does not necessarily imply that the vertical circulation model is one in which there is general ascending motion over a more or less deep layer of the troposphere on the high pressure side and subsidence on the low pressure side; it does, however, suggest that such a vertical circulation model exists farther upstream, probably near the jet stream entrance region as has been suggested by Namias and Clapp (5). Such a model of vertical motion would be consistent with horizontal divergence in the upper troposphere at about the level of the jet stream axis roughly balanced by convergence below in the lower troposphere on the high pressure side of the jet stream; high level convergence with low level divergence would then occur on the low pressure side of the jet stream. The association between extensive layer cloud and an active surface front can then be explained if the convergence-divergence field is pronounced, since the low level horizontal convergence, which is well known to have the effect of intensifying fronts, occurs on the high pressure side of the jet stream in the lower troposphere where the surface front is normally located.

It is interesting in this connection that flight 6 took place fairly near the jet stream entrance (see Fig. 5), and that in this case there was much layer cloud on the high pressure side of the axis as confirmed both by the aircraft and surface observations. None of the other flights was carried out in either a pronounced entrance or exit region.

## 7. Turbulence near jet streams

It is perhaps rather surprising that the twenty flights in the vicinity of jet streams experienced so little heavy turbulence. In about 26½ hours flying on the horizontal legs shown in Fig. 1, when observations were generally made every minute, only once was a bump of intensity 0.4 g (the lower limit of heavy turbulence according to Bannon (6)) reported. Some statistics of the occurrence of bumps, based on one report every minute, are shown in Table III.

Table III Percentage frequencies of occurrence of bumps of various intensities in the neighbourhood of 20 jet streams during 26½ hours flying on horizontal legs between 400 mb. and 200 mb.

| Intensity of turbulence | ≥ 0.1 g | ≥ 0.2 g | ≥ 0.3 g |
|---|---|---|---|
| Percentage frequency of occurrence | 37.4 | 4.9 | 0.2 |

Slight bumps, i.e. of intensity less than 0.2 g, occurred on roughly 33 per cent of the time, but heavy bumps were extremely rare; even bumps of moderate intensity occurred on less than 5 per cent of the time.

Bannon (6) has pointed out that there is a fairly high association between the occurrence of pronounced clear-air turbulence and small values of Richardson's Number (Ri), and also large horizontal wind shears. The present series of observations offers an opportunity to test Bannon's ideas on independent data.

The positions of the legs on which no bumps or bumps of intensity 0.1 g (i.e. slight bumpiness) were reported are shown in Fig. 16 in relation to the jet stream axis. It will be seen that there was some tendency for the legs with little or no bumpiness to be on the high pressure side of the jet stream axis. The minimum value of Ri (this suffers from the restriction that mean temperature lapse rates and vertical wind shears were measured over layers about 2000 feet thick) was estimated along each of the legs. The mean minimum Ri for all the legs was about 3.2, with individual minima ranging from about 0.5 to 7.6. Of the 14 legs, 12 had an estimated minimum Ri greater than unity. At the extremity, on the jet low pressure side, of each of the legs of flight 18 (synoptic situation shown in Fig. 4) Ri as measured was less than unity, but a displacement of the flight grid towards higher pressure by only 20 n.m. relative to the cross-section would have increased Ri to a value greater than unity. The positions of the legs relative to the jet streams are uncertain to about this extent since the flight and radio-sonde observations are not synchronous, and it is therefore possible that on this occasion Ri was greater than unity everywhere along the legs. It seems, therefore, that the lack of bumpiness on the legs shown in Fig. 16 is not inconsistent with Richardson's criterion that turbulence tends to increase with Ri less than unity.

Maximum values of horizontal wind shear were also estimated for the legs shown in Fig. 16. There is a good deal of uncertainty in estimating this quantity but not sufficient to mask the general results. Maximum cyclonic shear had a mean value about 25 kt. per 100 n.m. and varied from about 10 to 45 kt. per 100 n.m. in individual legs, maximum anticyclonic shear had a mean of about 21 kt. per 100 n.m. with individual variations from 10 to 38 kt. per 100 n.m. The two largest shears were cyclonic and occurred with flights 2 and 18. Flight 2 (see Fig. 6) was associated with a large Ri but flight 18, as has been mentioned, with Ri perhaps less than unity. In the latter case at least the horizontal shear might have been substantially less, for the same reason that Ri might have been larger than the computed value.

The nine regions (associated with five jet streams) with bumps greater than 0.2 g (i.e. 0.25 g to 0.4 g) are shown in Fig. 17. Broadly speaking, the bumpy positions marked in Fig. 17 are complementary to those with little bumpiness shown in Fig. 16, although there is some overlapping. It is particularly noteworthy that the sector in the troposphere on the high pressure side of the jet stream axis is completely free of turbulence of the scale under consideration. Bannon (7) has found a similar distribution of severe bumps. All the bumps above the jet stream axis occurred at or very near the tropopause. All but two groups occurred in clear air, including the most severe bumps (intensity 0.3 g to 0.4 g) reported on the series of flights, namely those of flight 7 at between 50 n.m. and 100 n.m. from the axis on the low pressure side; the other bumpy region (intensity 0.25 g) on flight 7 occurred in a cumulonimbus anvil, and the line of bumps (intensity 0.25 g) on flight 12 in nimbostratus cloud.

Values of Ri in the vicinity of the bumpy regions were computed for eight cases shown in Fig. 17 (the two adjacent bumpy regions of flight 7 were regarded as one). The mean value of Ri was about 1.2 with individual values ranging from about 0 to 4.2. The mean Ri in this case is significantly smaller than the mean minimum Ri estimated for the legs shown in Fig. 16. Furthermore in six out of the eight bumpy cases Ri was almost certainly less than unity. The isolated bump on the bottom leg of flight 12 and that on the top leg of flight 17 were probably associated with Ri greater than about three, but in each case the bump was only of intensity 0.25 g. However, it is noteworthy that the bump on flight 12 was associated with a strong horizontal cyclonic wind shear of about 80 kt. per 100 n.m., and that the bump on flight 17 occurred at the tropopause.

The horizontal wind shears in the neighbourhood of the bumpy regions shown in Fig. 17 were all cyclonic except for an anticyclonic shear of about 10-15 kt. per 100 n.m. in the case of the bump observed on flight 17. The cyclonic shears varied from at least about 20 kt. per 100 n.m. to 80 kt. per 100 n.m. from case to case with a mean value of about 40 kt. per 100 n.m.; but in some cases much stronger shears occurred quite near the estimated position, which is uncertain to some extent, of the bumpy region, so that it is possible that in fact the cyclonic shears varied from about 30 kt. to 90 kt. per 100 n.m. with a mean value of about 55 kt. per 100 n.m. Thus on the average the cyclonic shears were considerably larger for the bumpy cases (mean value between 40 and 55 kt. per 100 n.m. say) than for the non-bumpy cases (mean value about 25 kt. per 100 n.m.). It seems clear that there is some relationship between the occurrence of bumpiness and the strength of the horizontal shear, as has been suggested by Bannon (6). The present investigation would suggest as a rough working rule for forecasting purposes that turbulence of at least moderate intensity is probable when the horizontal wind shear is greater than about 45 kt. per 100 n.m., but may also occur with considerably lower values; shears of this intensity must almost certainly be cyclonic. With shears weaker than 45 kt. per 100 n.m. it is probable that bumps occur with small values of Ri as has been stated by Bannon (6).

It is interesting to observe that the most bumpy region reported during the flights occurred in clear air on the bottom leg of flight 7 below the jet stream axis, whereabouts Ri was small (less than unity) and the horizontal cyclonic shear was probably around 50 kt. per 100 n.m. Both conditions were suitable for the occurrence of turbulence. The rather singular case of the bump on flight 7 is difficult to explain; it occurred near the tropopause where the horizontal anticyclonic shear and the vertical shear were apparently quite weak.

## 8. Contrails

Contrails occurred on 13 legs (top leg of flights 6, 8, 9, 10, 11, 12, 13, 14, 16, 17, 18, 20 and bottom leg of flight 13), definitely did not occur on 2 legs (top leg of flight 19 and bottom leg of flight 14) and were not reported on the remaining 24 legs of the 20 special flights. It is possible that some ephemeral contrails formed and were not observed on these 24 legs, but it is at least fairly certain that no persistent contrails formed. On vertical climbs and descents observations were made rather intermittently, but it is unlikely that persistent contrails were missed.

The contrails invariably formed in regions where the air temperature was lower than the so-called Immunity-temperature given in M.O. 479 (8). However, there were many occasions when the observed temperature was lower than the Immunity-temperature but no contrails were reported. Short and ephemeral contrails were reported on several occasions in the lowest few thousand feet of the stratosphere: for example, on the top leg of flight 14, which was entirely in the stratosphere, non-persistent contrails occurred just above the tropopause (sloping) but ceased on that part of the leg which was more than about 3000 feet above the tropopause. The frost-point depressions in regions where contrails were observed varied very widely, but there was a broad difference between the non-persistent and very persistent contrails; the observed frost-point depressions ranged from 13 to 50°F. with non-persistent and from 1 to 26°F. with persistent contrails.

The observational facts given in the previous paragraph agree well with our knowledge of contrails which has recently been summarised in M.O. 479 (8). None of these features is peculiar to the atmosphere in the vicinity of jet streams.

## 9. Indicated-airspeed

The records of indicated-airspeed of the aircraft show fluctuations of the type that has been discussed by Frith (9), although it does not appear from inspection of the data in graphical form that there are any very significant correlations between these variations and changes of temperature or frost-point or bumpiness. Nor has it been found possible to relate the indicated-airspeed to synoptic features of the jet stream.

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## 10. Conclusions

(a) The special flights near jet streams confirm that the thermal distribution as obtained by means of cross-sections constructed in a fairly close network of routine aerological stations is substantially correct. Although the flight data, for various reasons, were not always adequate to give definite information on the structure of the tropopause on the low pressure side of the jet stream axis, they do not disagree with the drawing of a continuous tropopause on some and a disruption of the tropopause surface on other cross-sections, indicating that both types of structure appear to occur in nature. The temperature observations also confirm that small scale irregularities of about 1°F. occur in the lower stratosphere and near the tropopause, and significantly smaller fluctuations in the upper troposphere; but these are not specially features of the atmosphere near jet streams.

(b) There is on the average a horizontal gradient of humidity, in the same sense as the pressure gradient, across the jet stream in the region between about 500 mb. and 200 mb. The humidity of the air in the core of the jet stream varies considerably from case to case, but the average relative humidity with respect to ice at about the level of the jet stream is around 50 per cent at 300 mi. from the axis on the high pressure side and about 10 per cent at a similar distance on the low pressure side. Horizontal traverses of the tropopause demonstrate that the tropospheric air is moister than the stratospheric air at the same level, so that at least part of the horizontal humidity gradient may be accounted for by the existence of a sloping tropopause. There is also evidence for the existence of a relatively dry patch of air in the vicinity of the frontal zone below the jet stream axis; this dry feature of frontal zones has been noted by Sawyer (4).

(c) Cloud is variable from one jet stream to another, but there tends to be a broad pattern. The medium and high types of layer cloud predominate on the high pressure side of the jet stream axis, decrease in frequency of occurrence and in amount near the jet stream axis and are rare at a distance greater than about 100 n.m. from the jet stream axis on the low pressure side. Layer clouds do not appear to occur above the level of the jet stream axis, although occasionally patches of cirrus are observed in the uppermost part of the troposphere on the high pressure side. The jet streams which have fairly extensive layer clouds on the high pressure side are normally associated with surface fronts. Those jet streams with rather little layer cloud are generally associated with no surface front or a very weak one; such jet streams tend to be located on the forward side of an upper-ridge. The average distribution of layer cloud suggests a vertical circulation model with ascending motion in the troposphere on the high pressure side and descending motion on the low pressure side, operating upstream probably in the jet entrance region.

(d) On rather more than one third of the flying time near jet streams some turbulence was reported, but it was greater than slight in intensity on only about five per cent of the time. All but one of the moderate or severe bumps can be associated with small values of Richardson's Number (usually less than unity) or large horizontal wind shears (cyclonic) or both factors together, thus confirming on independent data what has been said by Bannon (6).

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/Appendix

## Appendix

List of flights carried out by the Meteorological Research Flight, Farnborough, from May 1951 to August 1952 in connection with the investigation of jet streams. Most of the flights were made in the afternoon, centred generally within an hour or so of the time of the routine aerological soundings; flights 10, 12, 16 and 20 took place some two to three hours earlier. Flight 15 was terminated after completion of bottom leg because of engine trouble. Instrumental failure resulted in the loss of observations on some flights: practically no temperatures were available on flight 5; frost-points were missing for this reason over much or all of the top legs of flights 1, 7, 11, 14 and 18 and over shorter distances on several other legs. The jet stream velocity given in the list below was obtained in each case from the wind maximum on the relevant cross-section.

| No. | Date | Bottom leg (mb.) | Top leg (mb.) | Jet velocity (kt.) | |
|---|---|---|---|---|---|
| 1 | May 3, 1951 | 400 | 206 | S. | 70 |
| 2 | May 17, 1951 | 325 | 275 | NE. | 100 |
| 3 | May 24, 1951 | 350 | 250 | S.-SSW. | 90 |
| 4 | May 30, 1951 | 350 | 250 | ENE. | 70 |
| 5 | June 15, 1951 | 350 | 250 | SW. | 90 |
| 6 | June 18, 1951 | 400 | 250 | SSW. | 130 |
| 7 | July 11, 1951 | 350 | 250 | SSW. | 110 |
| 8 | September 19, 1951 | 400 | 250 | NW.-NNW. | 100 |
| 9 | December 3, 1951 | 400 | 250 | NNW.-N. | 110 |
| 10 | December 10, 1951 | 400 | 250 | NW.-NNW. | 130 |
| 11 | December 10, 1951 | 400 | 250 | NW.-NNW. | 130 |
| 12 | January 2, 1952 | 400 | 250 | W. | 130 |
| 13 | January 16, 1952 | 350 | 200 | W. | 90 |
| 14 | February 1, 1952 | 350 | 250 | WNW.-NW. | 150 |
| 15 | February 5, 1952 | 400 | - | N. | 140 |
| 16 | March 21, 1952 | 350 | 250 | NW.-NNW. | 120 |
| 17 | March 26, 1952 | 400 | 220 | NNW. | 80 |
| 18 | April 24, 1952 | 400 | 250 | NNW. | 90 |
| 19 | August 12, 1952 | 400 | 250 | SW. | 110 |
| 20 | August 13, 1952 | 400 | 250 | SW. | 110 |

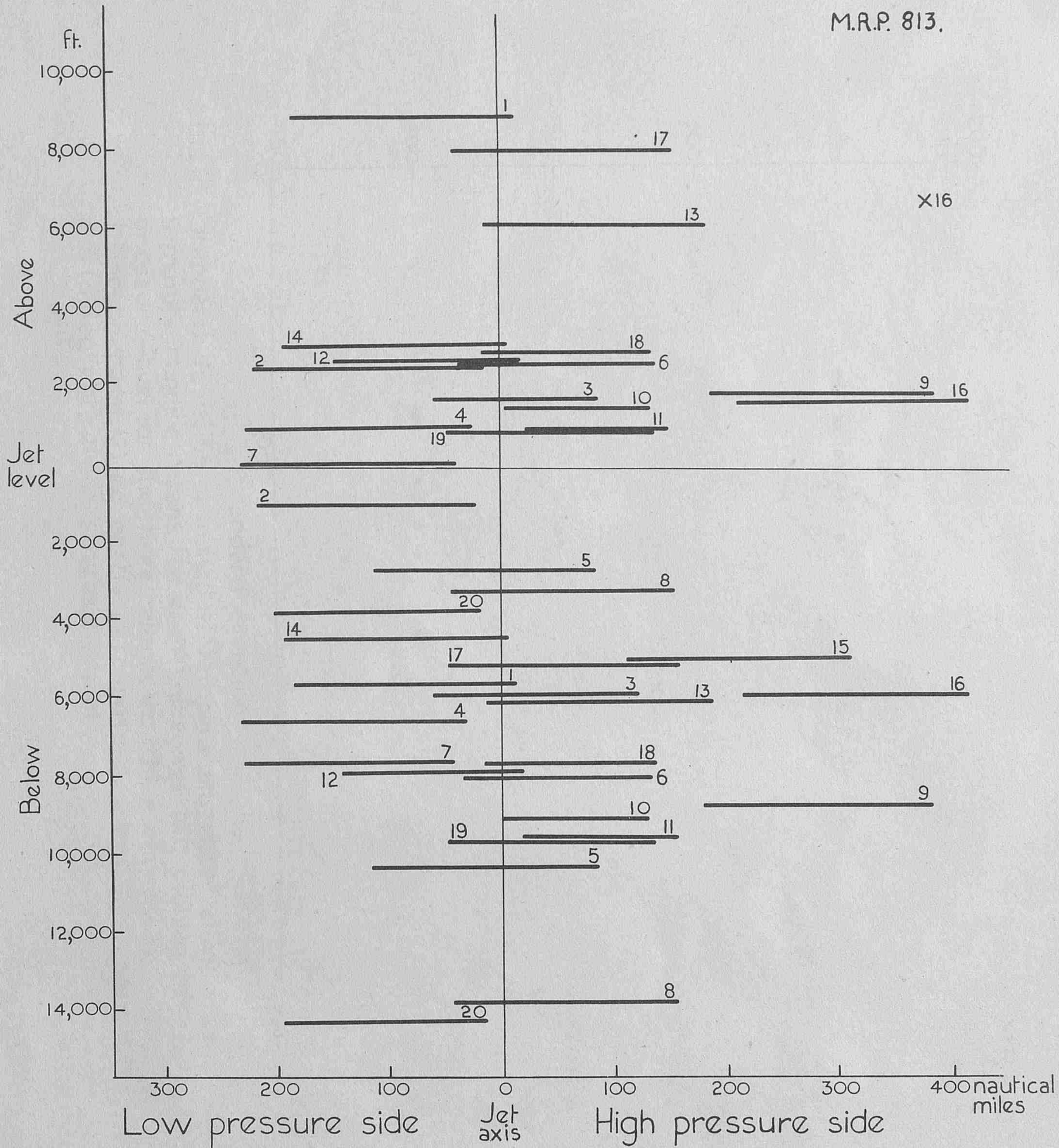

Fig.1 Positions of horizontal legs of flights relative to jet stream axis. With flight 16, a minor jet stream centre is marked thus: X16

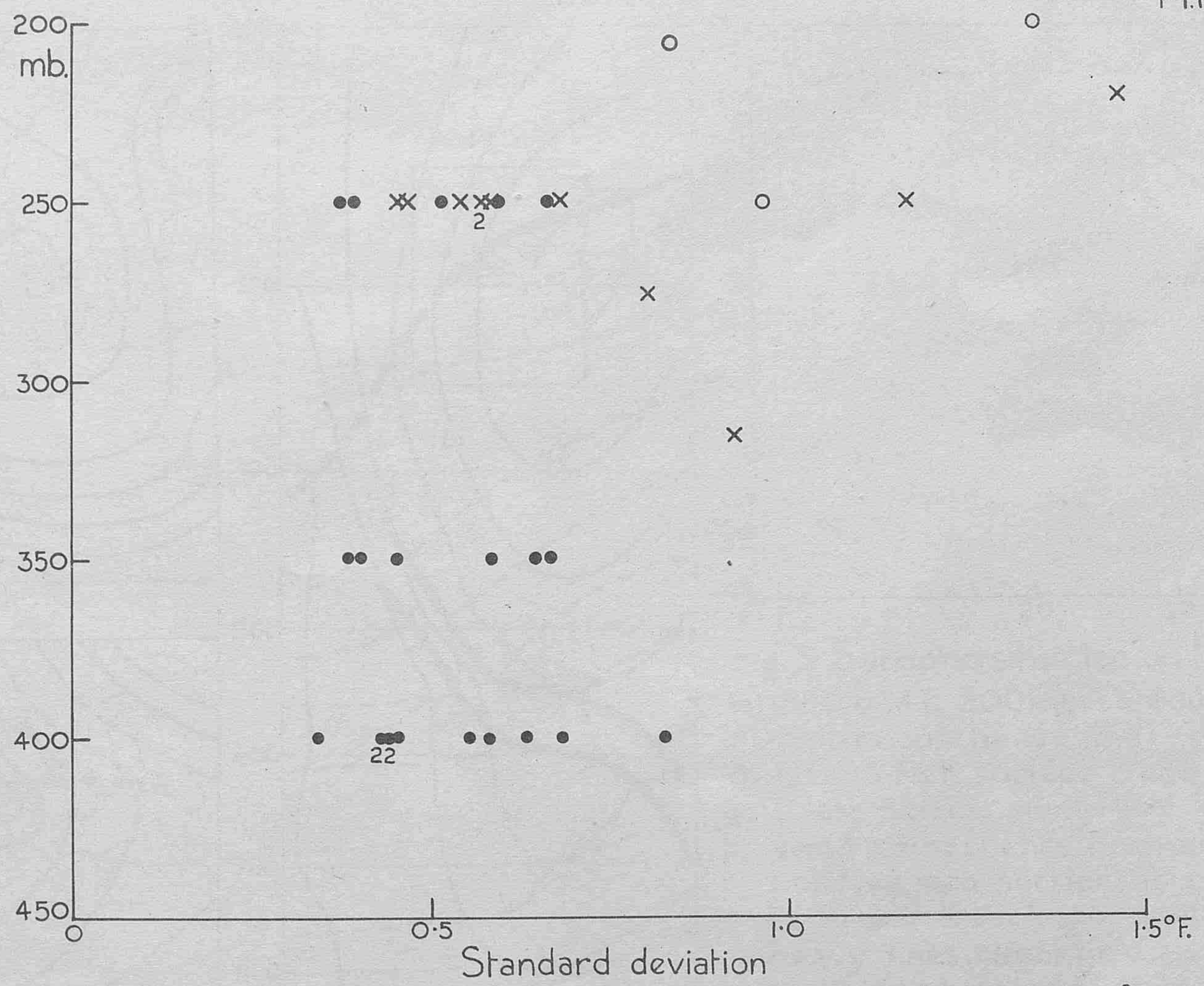

Fig.2 Standard deviation of individual temperature observations from 9-minute running means for each horizontal leg, plotted against pressure-level for the three categories, (1) leg entirely in troposphere (•), (2) leg partly in troposphere and stratosphere (x) and (3) leg entirely in stratosphere (o)

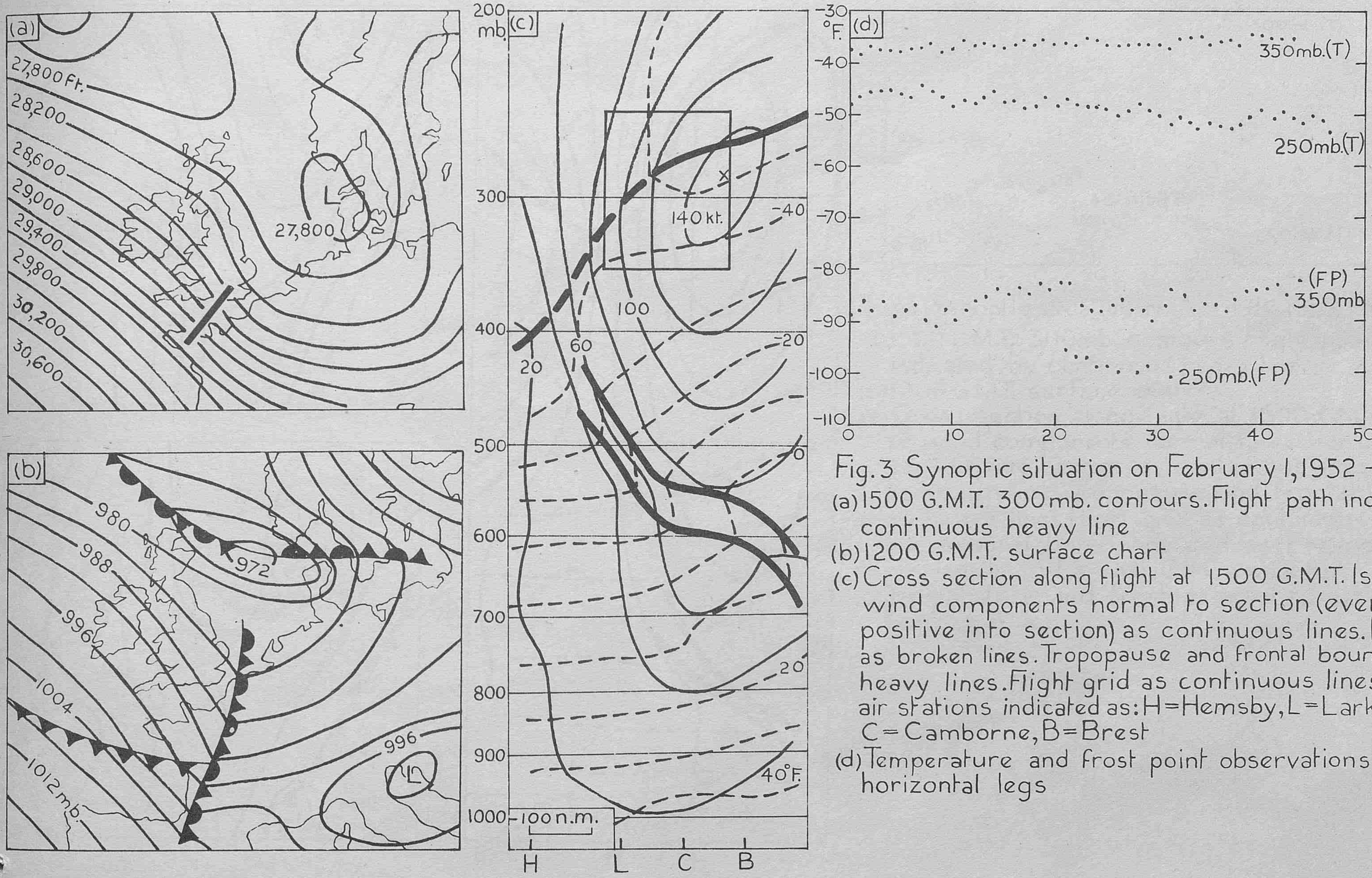

Fig. 3 Synoptic situation on February 1, 1952 — flight 14

(a) 1500 G.M.T. 300 mb. contours. Flight path indicated by continuous heavy line

(b) 1200 G.M.T. surface chart

(c) Cross section along flight at 1500 G.M.T. Isopleths of wind components normal to section (every 20 kt., positive into section) as continuous lines. Isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Flight grid as continuous lines. Upper air stations indicated as: H = Hemsby, L = Larkhill, C = Camborne, B = Brest

(d) Temperature and frost point observations on horizontal legs

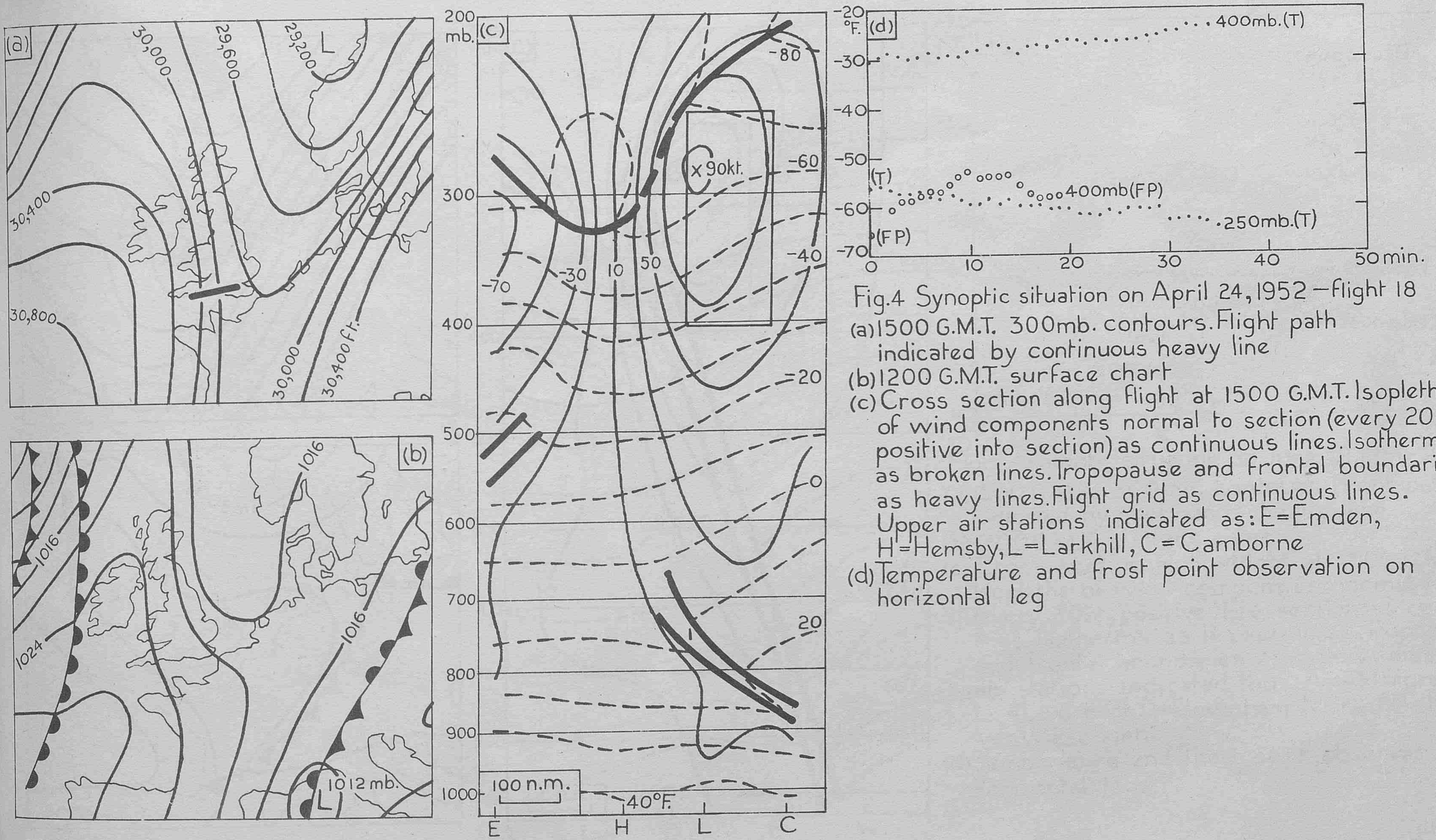

Fig.4 Synoptic situation on April 24, 1952 – flight 18

(a) 1500 G.M.T. 300mb. contours. Flight path indicated by continuous heavy line

(b) 1200 G.M.T. surface chart

(c) Cross section along flight at 1500 G.M.T. Isopleths of wind components normal to section (every 20kt., positive into section) as continuous lines. Isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Flight grid as continuous lines. Upper air stations indicated as: E=Emden, H=Hemsby, L=Larkhill, C=Camborne

(d) Temperature and frost point observation on horizontal leg

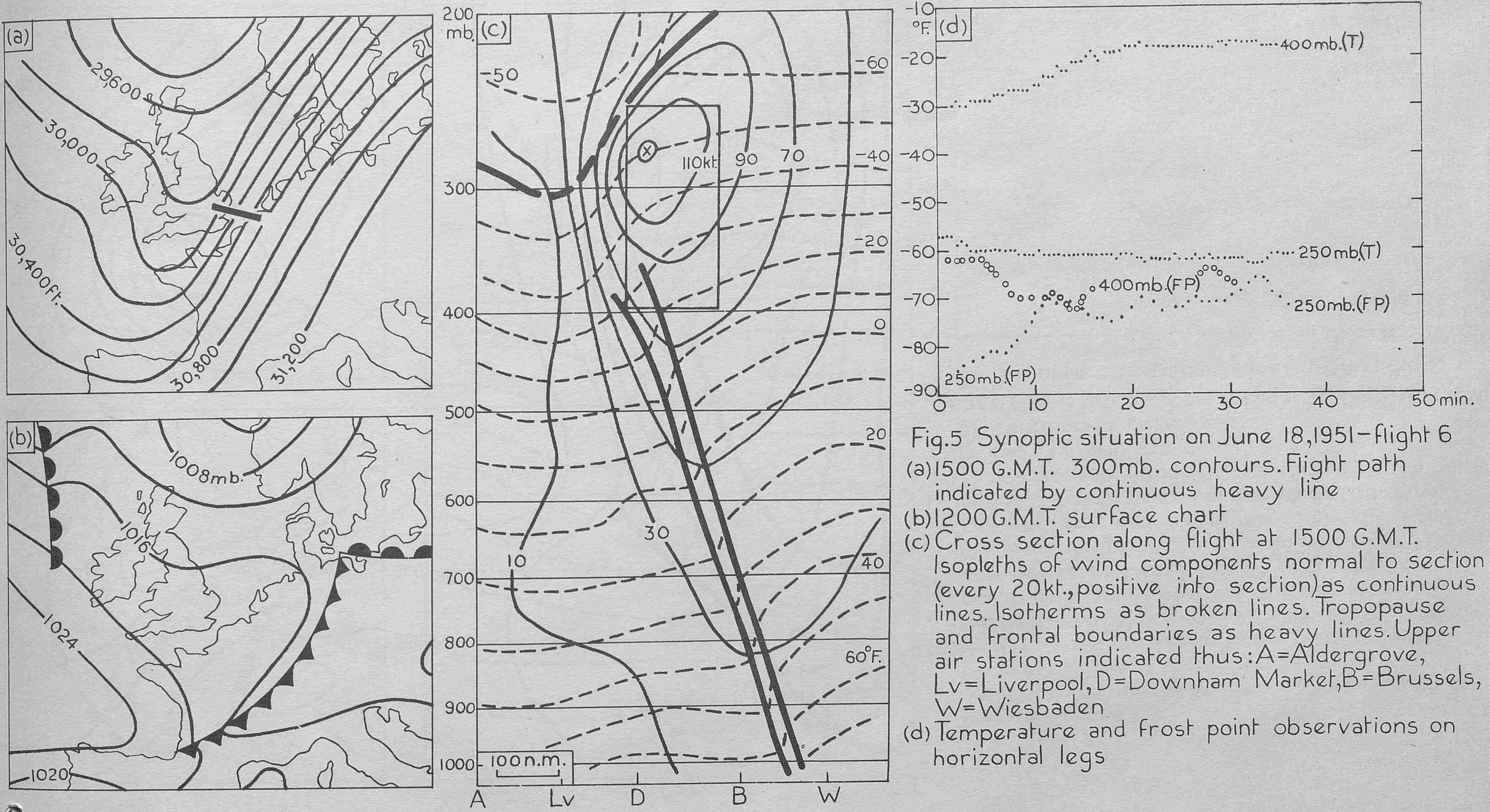

Fig.5 Synoptic situation on June 18,1951–flight 6

(a) 1500 G.M.T. 300mb. contours. Flight path indicated by continuous heavy line

(b) 1200 G.M.T. surface chart

(c) Cross section along flight at 1500 G.M.T. Isopleths of wind components normal to section (every 20kt., positive into section) as continuous lines. Isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Upper air stations indicated thus: A=Aldergrove, Lv=Liverpool, D=Downham Market, B=Brussels, W=Wiesbaden

(d) Temperature and frost point observations on horizontal legs

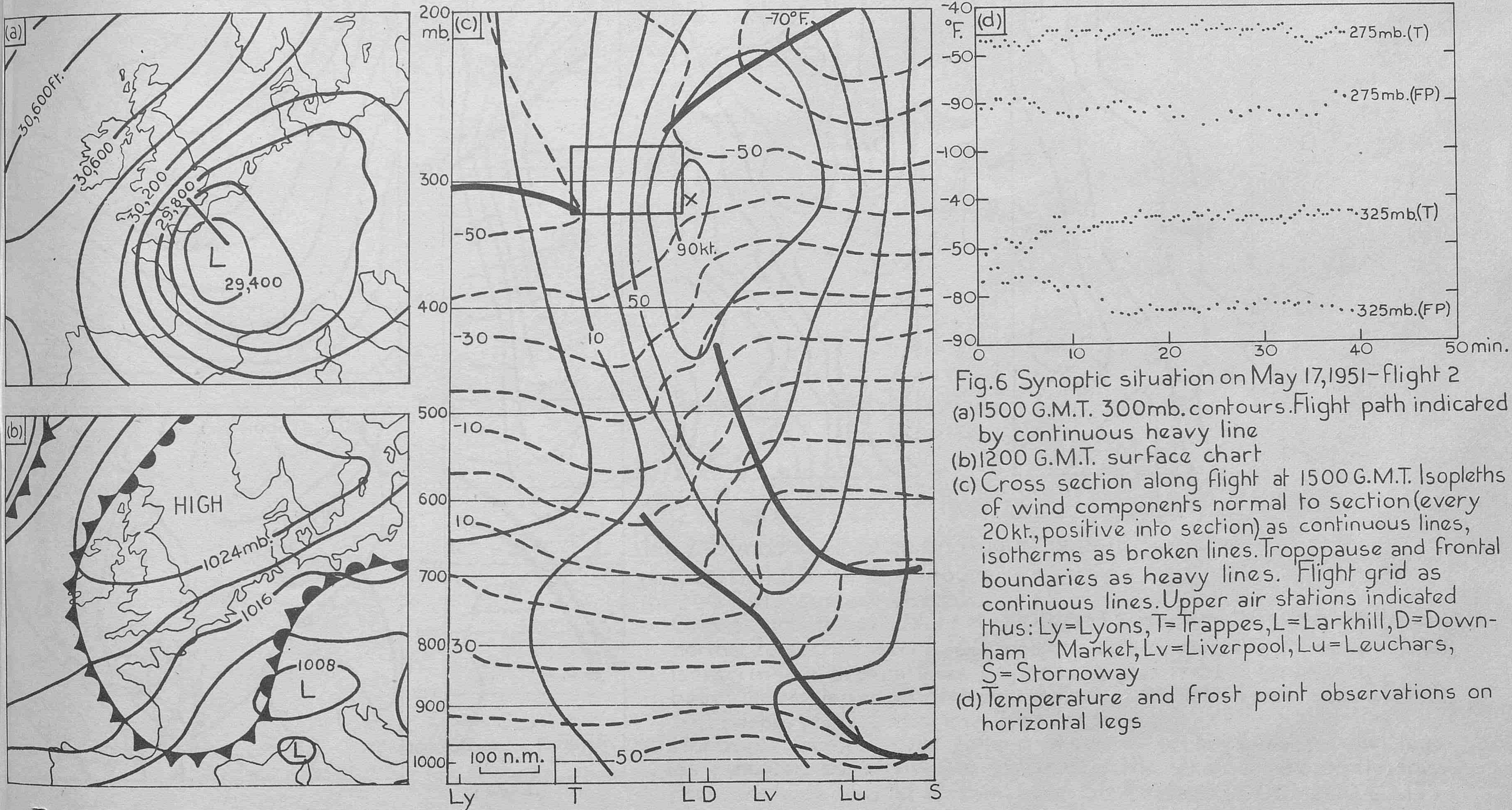

Fig. 6 Synoptic situation on May 17, 1951 – flight 2

(a) 1500 G.M.T. 300 mb. contours. Flight path indicated by continuous heavy line

(b) 1200 G.M.T. surface chart

(c) Cross section along flight at 1500 G.M.T. Isopleths of wind components normal to section (every 20 kt., positive into section) as continuous lines, isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Flight grid as continuous lines. Upper air stations indicated thus: Ly = Lyons, T = Trappes, L = Larkhill, D = Downham Market, Lv = Liverpool, Lu = Leuchars, S = Stornoway

(d) Temperature and frost point observations on horizontal legs

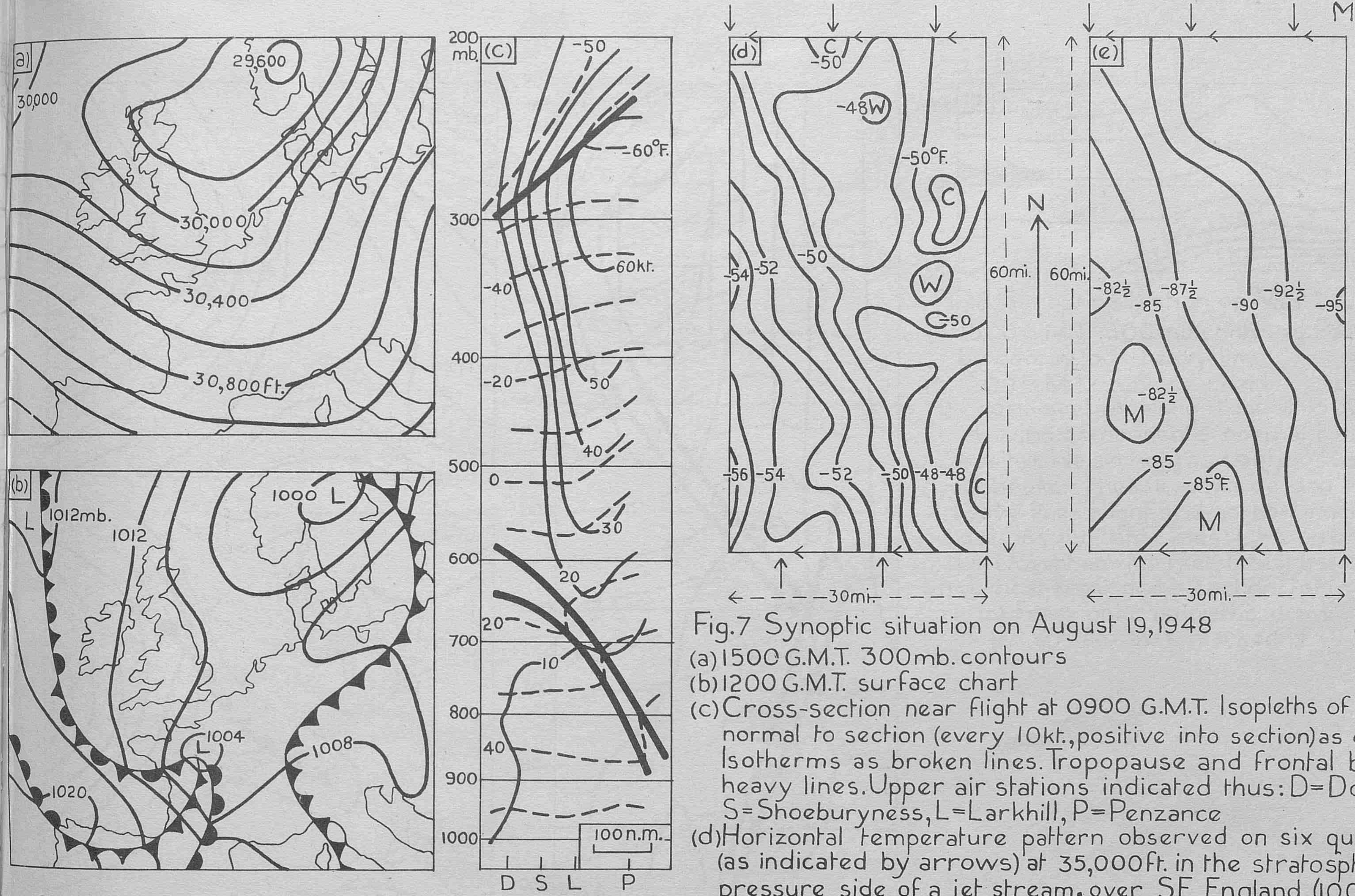
(a)
30,000
29,600
30,000
30,400
30,800 ft.
(b)
L
1012 mb.
1012
1000 L
L
1004
1008
1020
(c)
200 mb.
300
400
500
600
700
800
900
1000
-50
-60°F.
60 kt.
-40
-20
50
40
0
30
20
20
10
40
100 n.m.
D S L P
(d)
C
-50
-48 W
-50°F.
C
W
-50
-54 -52
-50
-56 -54
-52
-50 -48 -48
C
60 mi.
N
30 mi.
(e)
60 mi.
-82½
-85
-87½
-90
-92½
-95
M
-82½
-85
-85°F.
M
30 mi.

Fig.7 Synoptic situation on August 19, 1948

(a) 1500 G.M.T. 300 mb. contours

(b) 1200 G.M.T. surface chart

(c) Cross-section near flight at 0900 G.M.T. Isopleths of wind components normal to section (every 10 kt., positive into section) as continuous lines. Isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Upper air stations indicated thus: D = Downham Market, S = Shoeburyness, L = Larkhill, P = Penzance

(d) Horizontal temperature pattern observed on six quasi-parallel legs (as indicated by arrows) at 35,000 ft. in the stratosphere, on the low pressure side of a jet stream, over SE England. (1000-1200 G.M.T.) C and W indicate relatively cold and warm patches

(e) Horizontal frost point pattern associated with temperature pattern (d). Isopleths drawn every 2½°F. M indicates relatively moist patches

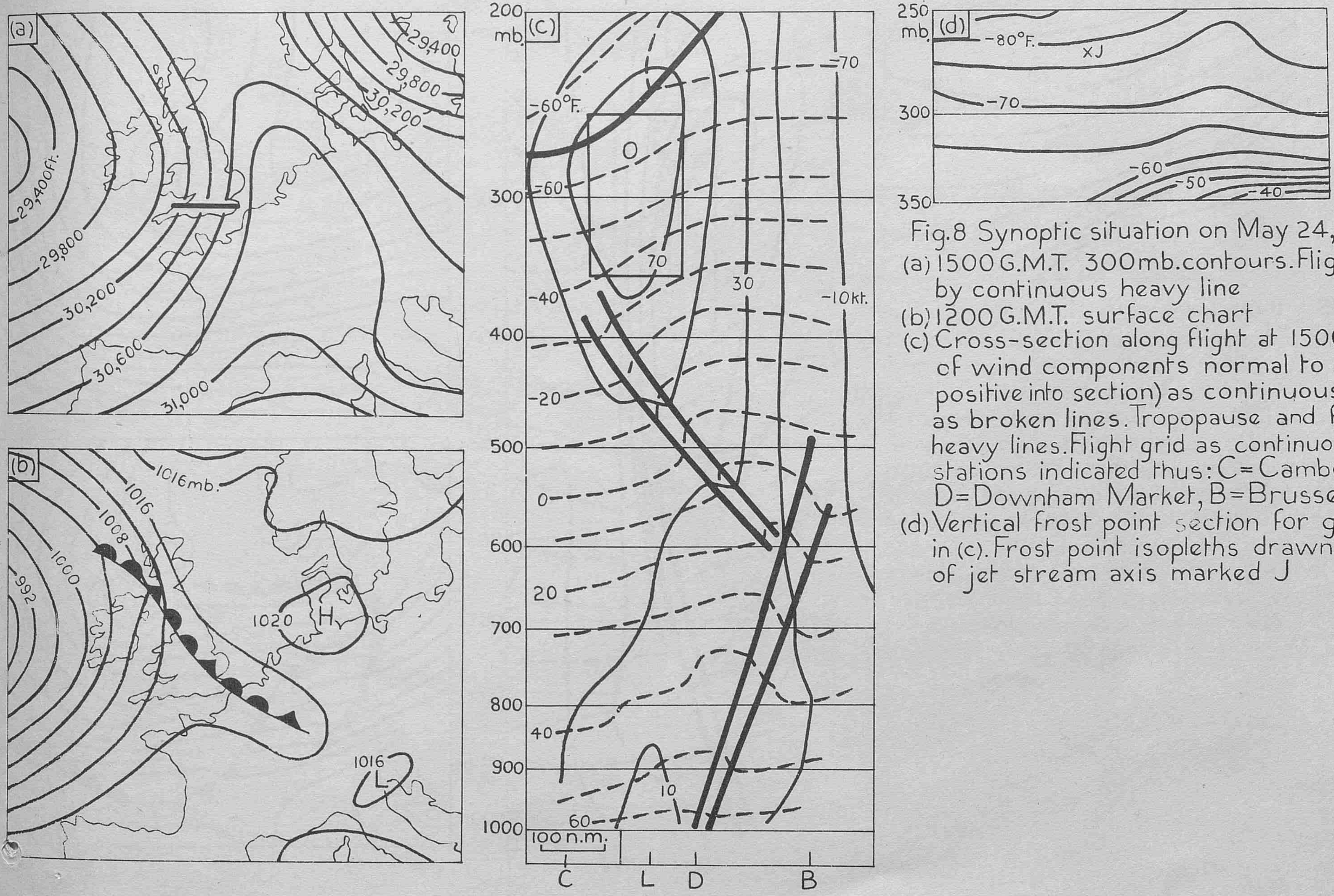

Fig.8 Synoptic situation on May 24, 1951 — flight 3

(a) 1500 G.M.T. 300 mb. contours. Flight path indicated by continuous heavy line

(b) 1200 G.M.T. surface chart

(c) Cross-section along flight at 1500 G.M.T. Isopleths of wind components normal to section (every 20 kt., positive into section) as continuous lines. Isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Flight grid as continuous lines. Upper air stations indicated thus: C = Camborne, L = Larkhill, D = Downham Market, B = Brussels

(d) Vertical frost point section for grid region shown in (c). Frost point isopleths drawn every 5°F. Position of jet stream axis marked J

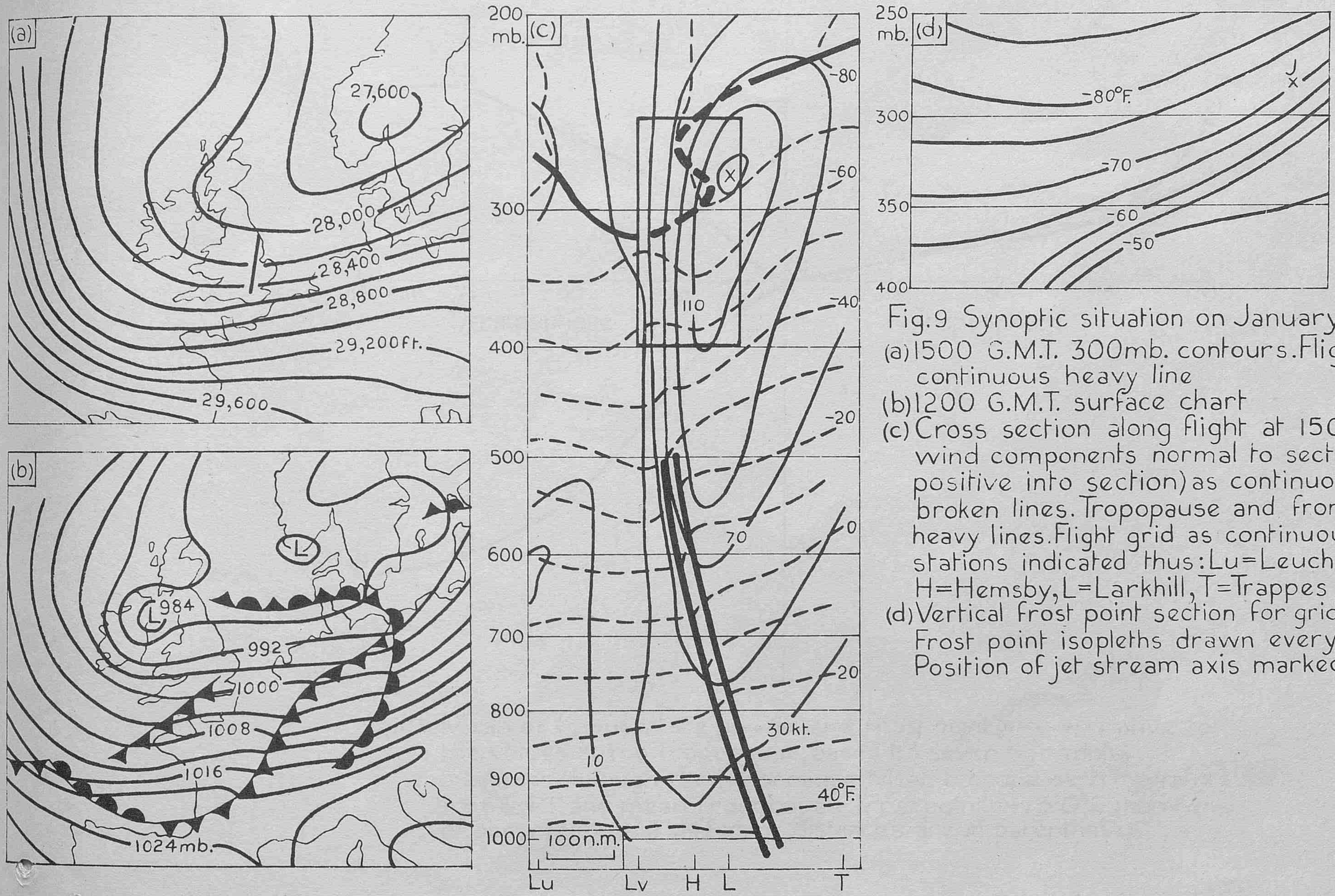

Fig. 9 Synoptic situation on January 2, 1952 — flight 12

(a) 1500 G.M.T. 300mb. contours. Flight path indicated by continuous heavy line

(b) 1200 G.M.T. surface chart

(c) Cross section along flight at 1500 G.M.T. Isopleths of wind components normal to section (every 20 kt., positive into section) as continuous lines. Isotherms as broken lines. Tropopause and frontal boundaries as heavy lines. Flight grid as continuous lines. Upper air stations indicated thus: Lu=Leuchars, Lv=Liverpool, H=Hemsby, L=Larkhill, T=Trappes

(d) Vertical frost point section for grid region shown in (c). Frost point isopleths drawn every 5°F. Position of jet stream axis marked J

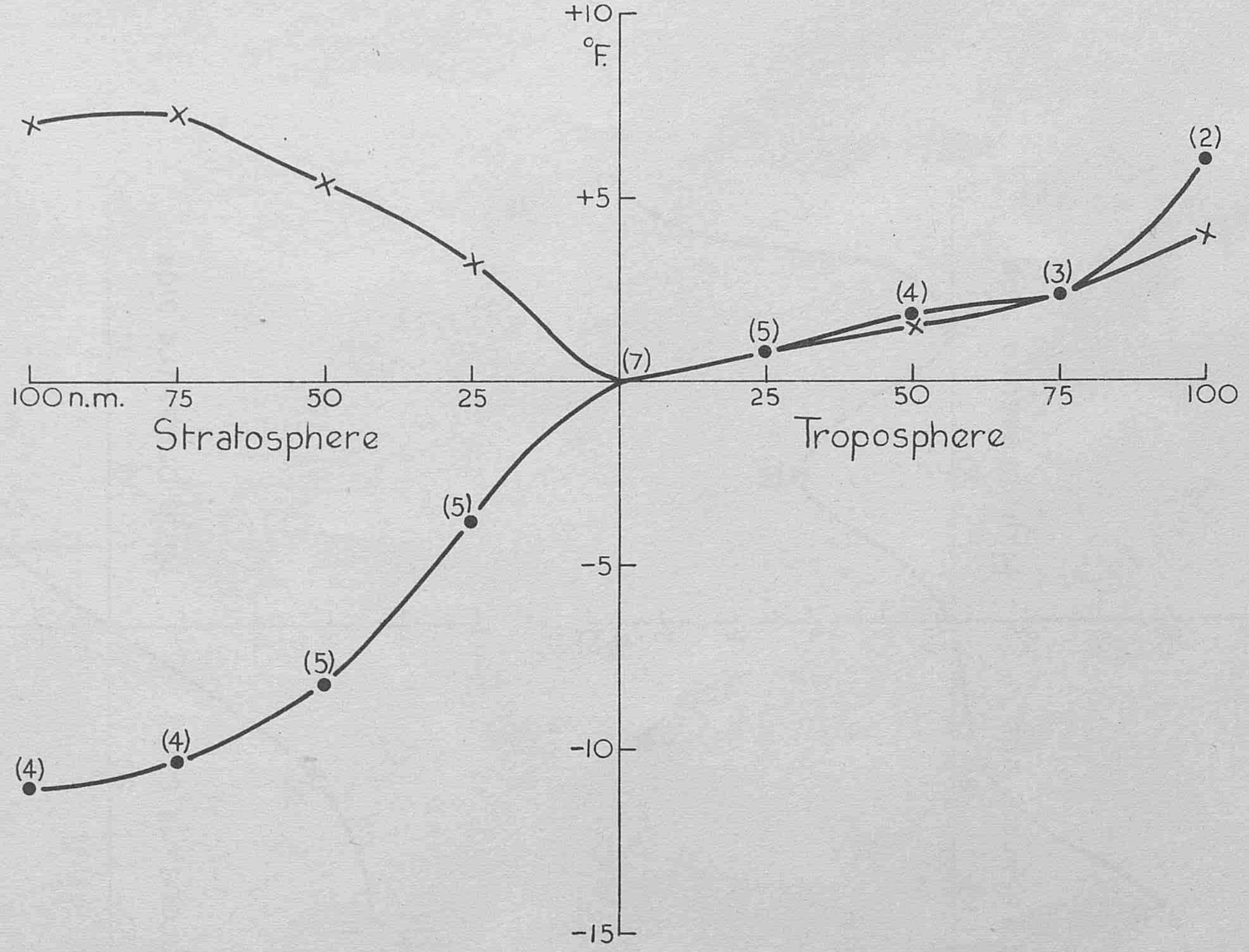

Fig.10 Mean of temperature (x——x) and frost point (●——●) relative to the values at the tropopause, based on seven horizontal traverses through the tropopause. Mean tropopause temperature = −64·4°F. and mean tropopause frost point = −75·0°F. : numbers of observations at different distances are in parentheses

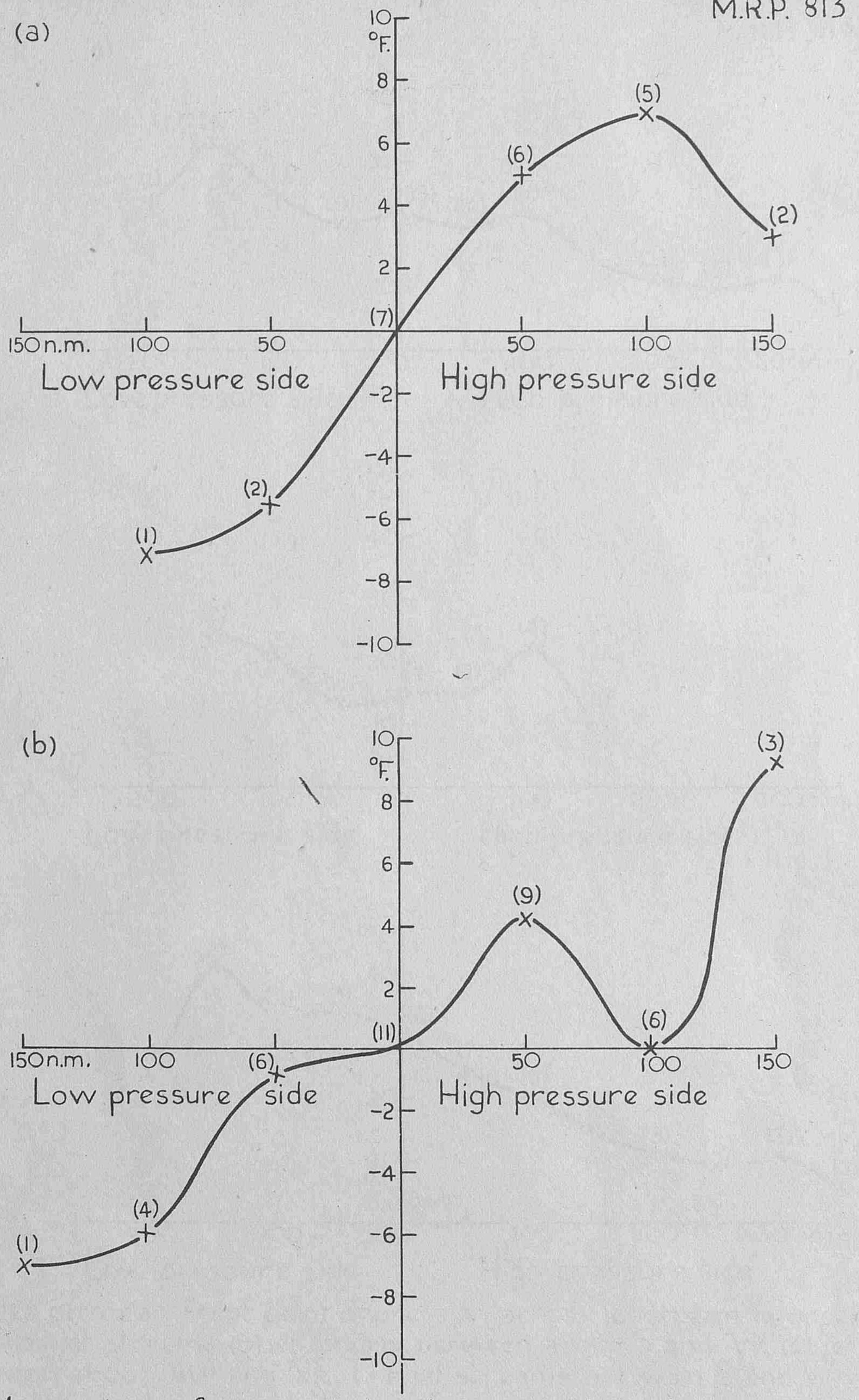

Fig.11 Mean relative frost point distribution (a) above jet stream axis and (b) below jet stream axis. Relative frost point is frost point minus that at point of intersection of horizontal leg with vertical through jet stream axis. Number of observations comprising mean values at various distances indicated in parentheses

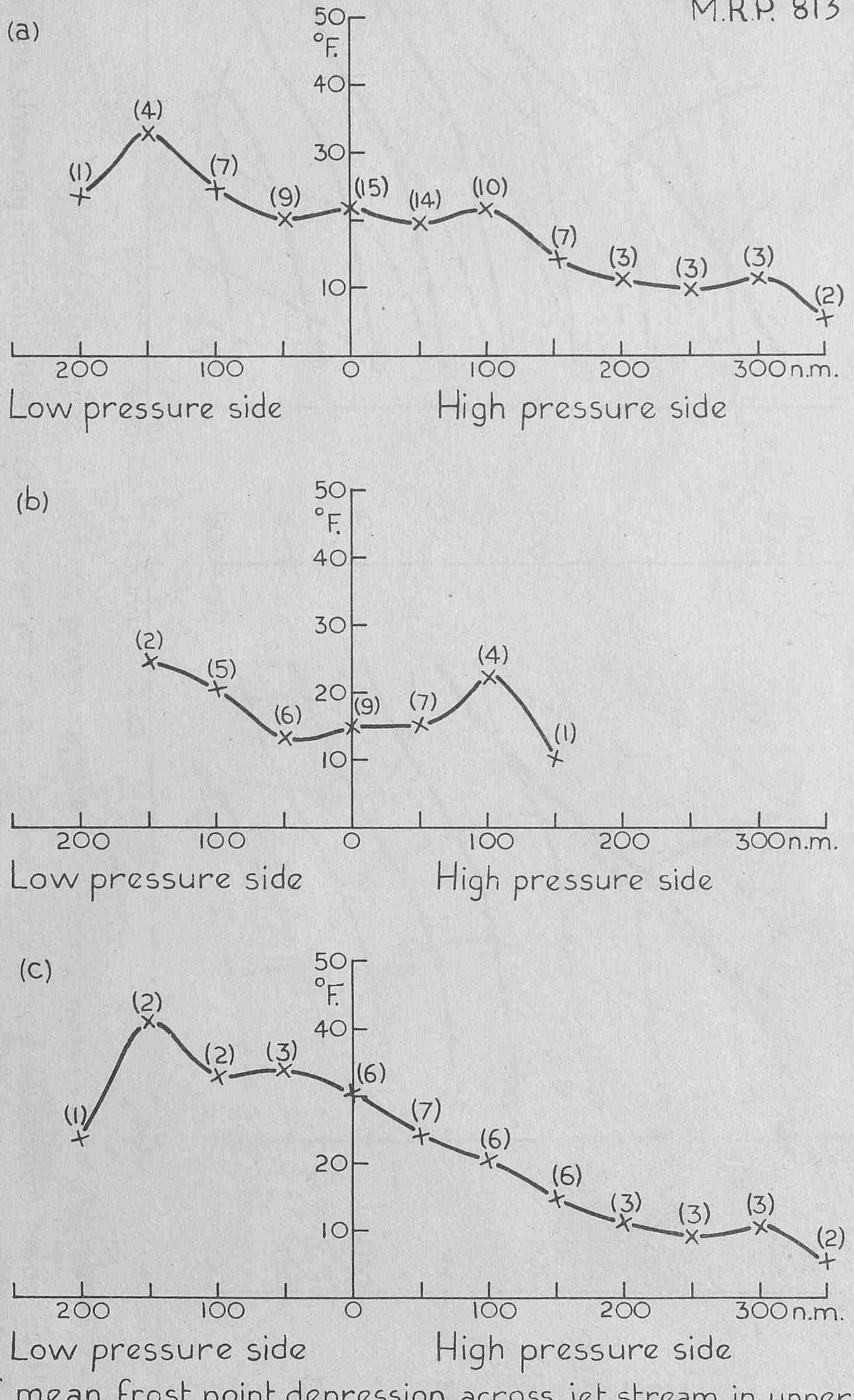

Fig. 12 Profiles of mean frost point depression across jet stream in upper troposphere for (a) all jet streams, (b) jet streams between about S and W, (c) jet streams between about NW and NE. The jet streams between S and W were either quasi-stationary on forward side of a cold upper trough or moved laterally towards the warmer tropospheric air. The jet streams between NW and NE were either quasi-stationary on forward side of a warm upper ridge or moved laterally towards the colder tropospheric air. Numbers of observations at various distances from jet stream axis are shown in parentheses

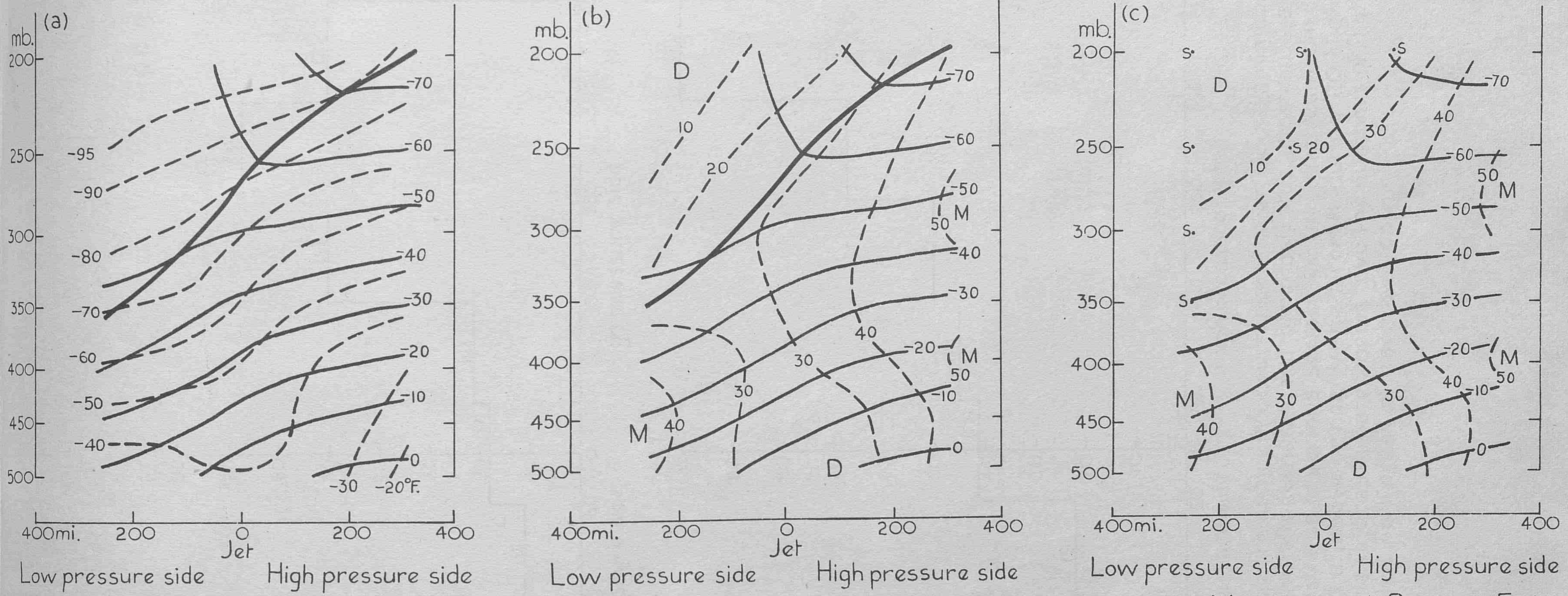

Fig.13 Mean sections showing humidity distribution near jet stream based on vertical soundings made by the Meteorological Research Flight during 1949-1951: (a) mean frost point and temperature distribution when no allowance is made for the fact that mean values at some points are based on both stratospheric and tropospheric observations; (b) as (a) but showing relative humidity w.r.t. ice instead of frost point; (c) mean relative humidity w.r.t. ice and temperature based on mean values entirely in the stratosphere (marked S) and in the troposphere. Mean tropopause as thick line; temperature and frost point in °F.; relatively dry and moist regions marked with D and M respectively

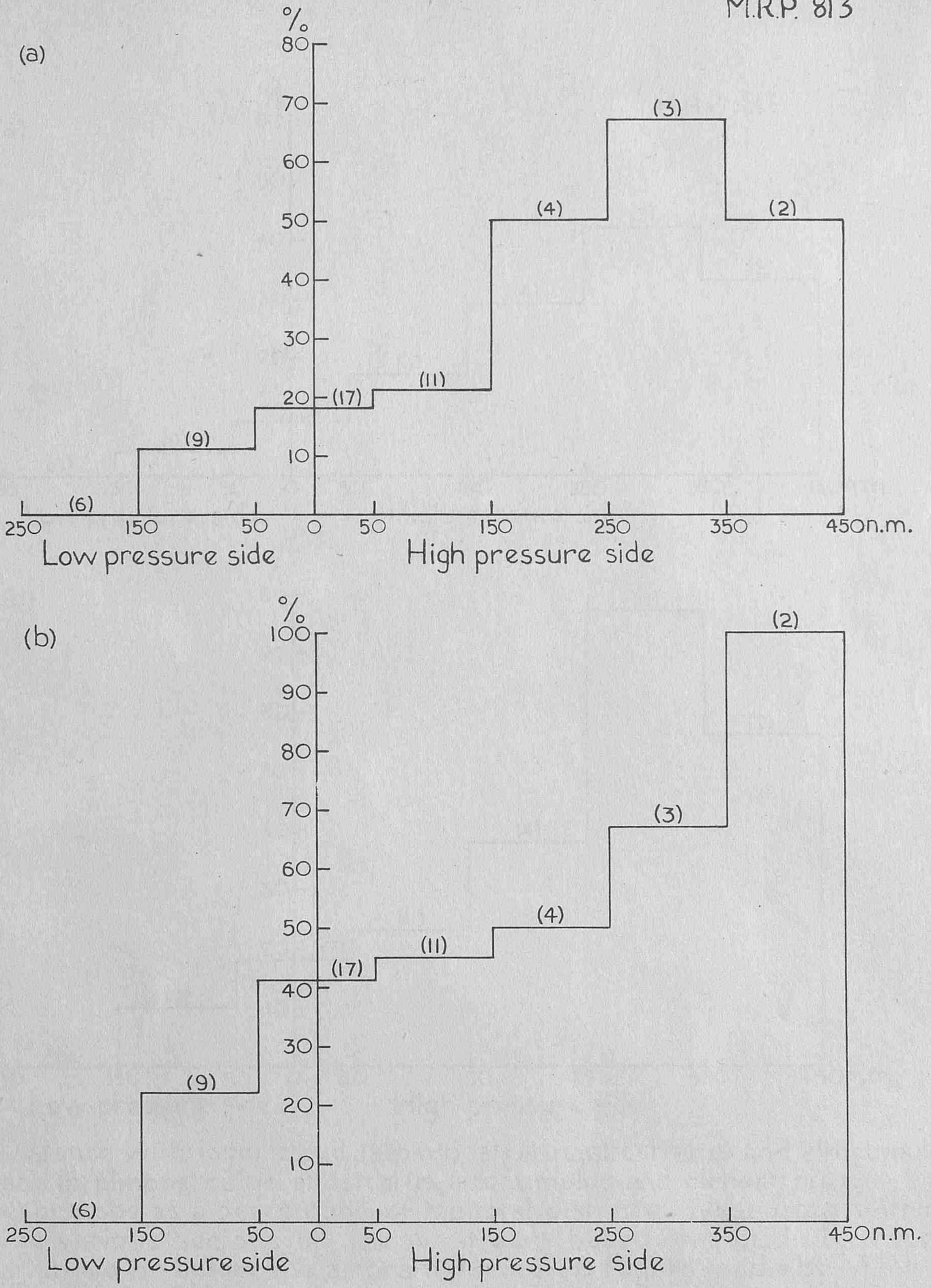

Fig. 14 Percentage frequency of occurrence of layer cloud, (a) nimbostratus, altostratus and altocumulus, and (b) nimbstratus, altostratus, altocumulus and cirrostratus in various sectors of the jet stream, based on cloud observed at, above or below the aircraft during 20 flights made by M.R.F. The figures in parentheses give the numbers of flights which investigated part or all of each sector

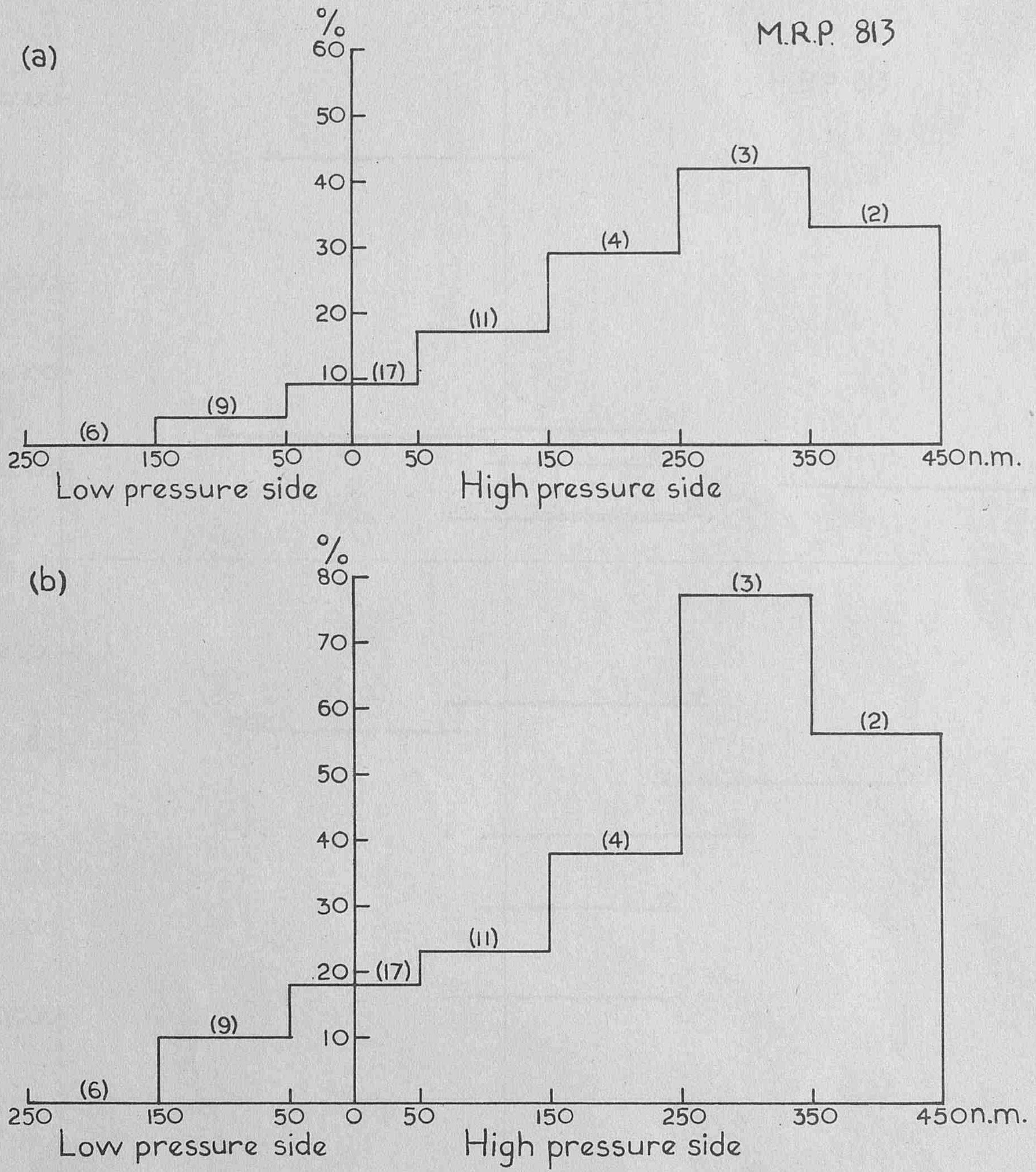

Fig.15 Distance with layer cloud,(a) nimbostratus, altostratus and altocumulus and (b) nimbostratus, altostratus, altocumulus and cirrostratus, expressed as a percentage of the total distance investigated within the various sectors of the jet stream, based on cloud observed at, above or below the aircraft during 20 flights made by M.R.F. The figures in parentheses give the number of flights which investigated part or all of each sector

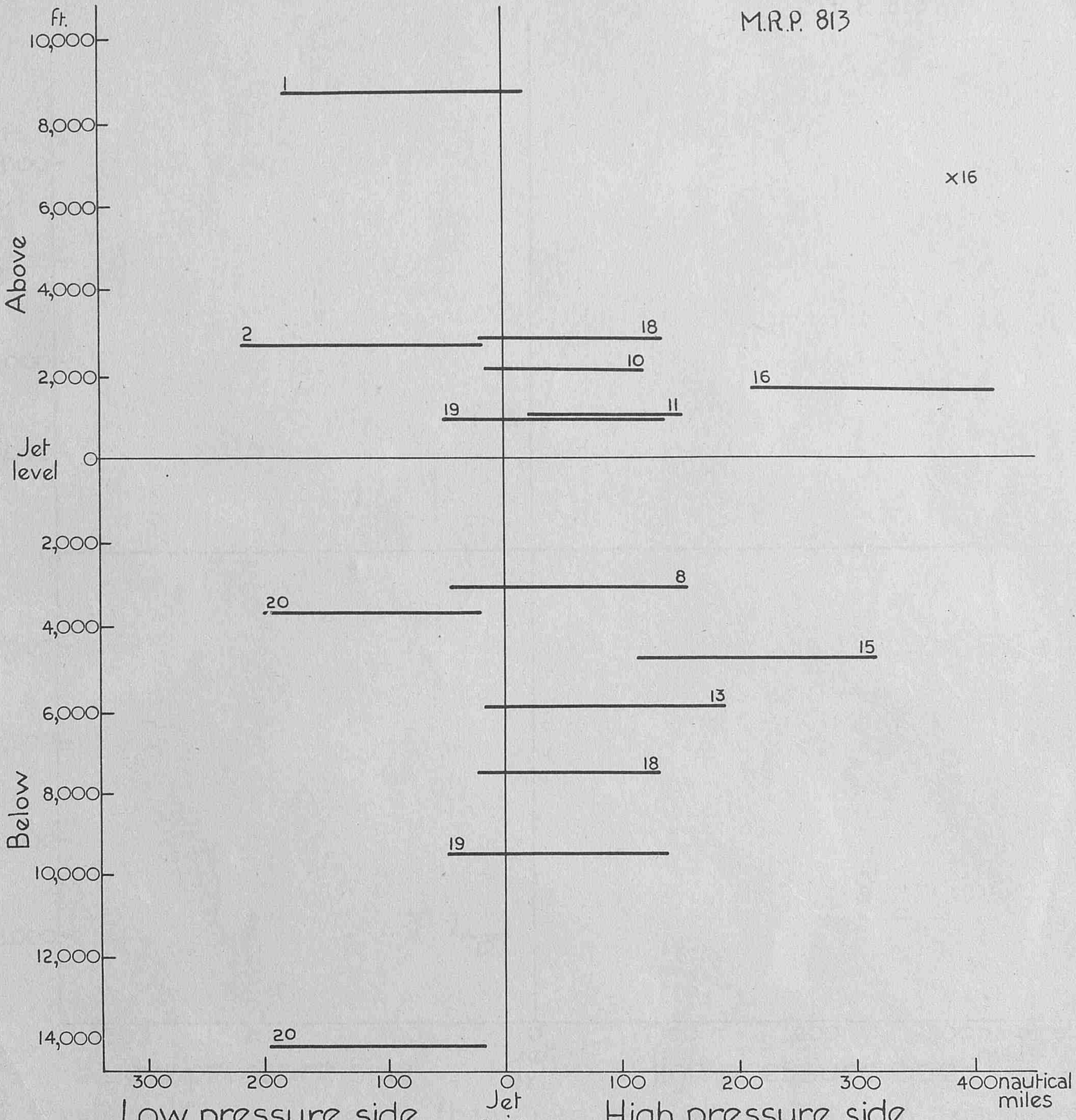

Fig.16 Positions of legs relative to jet stream axis with little or no bumpiness, i.e. with bumps not greater than 0·1g. The figures against each leg refer to the flights concerned, and a secondary jet stream centre is indicated with a cross

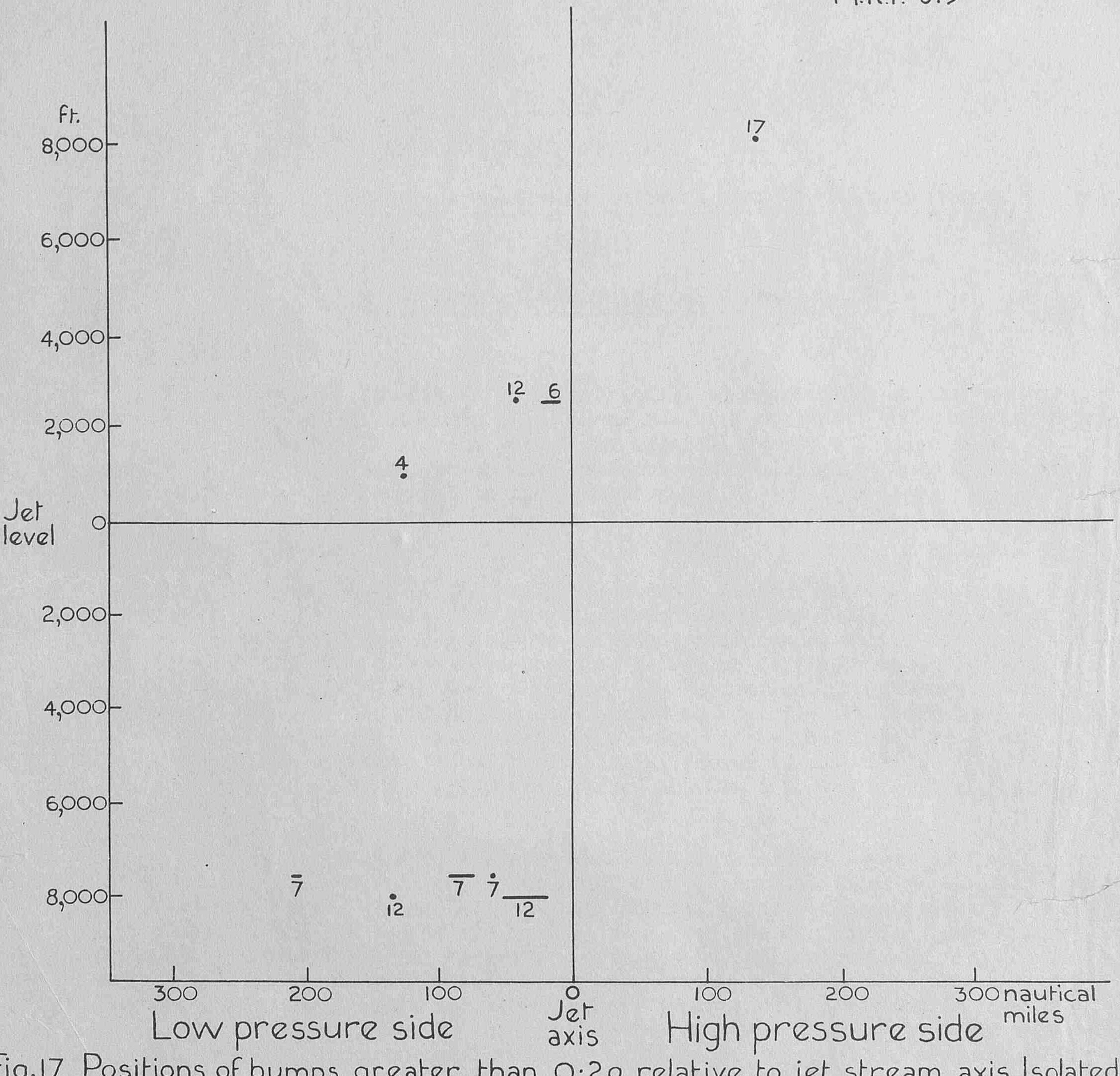

Fig. 17 Positions of bumps greater than 0·2g relative to jet stream axis. Isolated bump shown as a dot, line of bumps as a short line; flight number shown beside bumpy region

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S.C. II/153
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# AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# Rainfall of depressions which move eastward near the English Channel

by

J.S. Sawyer (Meteorological Office)

### 1. Introduction

In two previous investigations the rainfall of depressions moving eastward across England and to the north of Scotland has been examined, and the results have been reported in M.R.Ps. 748 and 762. The rainfall over the British Isles resulting from depressions which move eastward near the English Channel has now been studied and is discussed in the present paper.

### 2. Method

From an examination of the Daily and Monthly Weather Reports all occasions were noted on which a depression moved eastward along the English Channel or over adjacent coastal areas. In order to provide a homogeneous set of examples, it was required that there should be a closed isobar on at least one of the charts published in the D.W.R. and that the track of the depression between 5°W and 1°E should lie entirely within the area shaded in Fig. 1. In order to eliminate some depressions moving southeastward into France it was also required that the track should lie north of the 48°N parallel as far as longitude 5°E. [illegible] depressions were found in the 10 year period 1941 to 1950 which satisfied these conditions.

The methods of analysis followed those of the previous investigations. For each depression a rainfall chart was drawn for the British Isles on the basis of observations printed in the Daily Weather Report. The period covered by the chart was chosen to include all of the frontal rain from the depression; showers in the adjacent air masses were necessarily included, and difficulty in separating the rain from that of the preceding or following systems arose in some cases. However, the general shape and intensity of the rainfall distribution were rarely in doubt. An examination of the synoptic reports from selected stations was made in order to ascertain the northern limit of the area affected by the continuous rain-belt, and this was indicated on the rainfall chart. Most of the rainfall charts covered a period of 36 hours or less, but in six cases it was necessary to include a longer period; in two cases this extended to 60 hours.

For each depression a 1000-500 mb thickness chart was constructed; this was drawn for a time chosen so that the chart should represent the thermal structure of the depression as it crossed England. Observations were taken from the German Daily Weather Reports to complete the British charts for the war period.

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M.R.P. 816
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27th May, 1953

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### 2. Method

From an examination of the Daily and Monthly Weather Reports all occasions were noted on which a depression moved eastward along the English Channel or over adjacent coastal areas. In order to provide a homogeneous set of examples, it was required that there should be a closed isobar on at least one of the charts published in the D.W.R. and that the track of the depression between 5°W and 1°E should lie entirely within the area shaded in Fig. 1.; in order to eliminate some depressions moving southeastward into France it was also required that the track should lie north of the 48°N parallel as far as longitude 5°E. 42 depressions were found in the 10 year period 1941 to 1950 which satisfied these conditions.

The methods of analysis followed those of the previous investigations. For each depression a rainfall chart was drawn for the British Isles on the basis of observations printed in the Daily Weather Report. The period covered by the chart was chosen to include all of the frontal rain from the depression; showers in the adjacent air masses were necessarily included, and difficulty in separating the rain from that of the preceding or following systems arose in some cases. However, the general shape and intensity of the rainfall distribution were rarely in doubt. An examination of the synoptic reports from selected stations was made in order to ascertain the northern limit of the area affected by the continuous rain-belt, and this was indicated on the rainfall chart. Most of the rainfall charts covered a period of 36 hours or less, but in six cases it was necessary to include a longer period; in two cases this extended to 60 hours.

For each depression a 1000-500 mb thickness chart was constructed; this was drawn for a time chosen so that the chart should represent the thermal structure of the depression as it crossed England. Observations were taken from the German Daily Weather Reports to complete the British charts for the war period.

### 3. The average rainfall distribution

Fig. 2 shows the average rainfall distribution for the periods during which the depressions passed eastward near the English Channel. It is based on the average rainfall values shown on it and, in calculating these, values have been

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interpolated from the rainfall charts when observations were not available in the Daily Weather Report. Since the stations for which observations are included in the Daily Weather Report are nearly all on relatively low ground the chart may not adequately represent the rainfall of the hilly areas. However orographic influences on the rainfall of depressions on this track do not appear to be strongly developed.

In order to provide some indication of the rainfall in such notably wet areas as Dartmoor and the Welsh mountains, the average rainfall has been included as an inset in Fig. 2 for Princetown, Cray Reservoir (Breconshire) and Cwm Dyli near Snowdon. So far as possible these values are based on autographic raingauge records, but for Princetown values in the earlier years have had to be estimated from rainfall measurements made once daily. Average rainfalls for Eskdalemuir and Dalwhinnie, high level stations in Scotland, are also included in the inset to Fig. 2. These are based on observations published in the Daily Weather Report.

Fig. 2 shows that the average rainfall falls off steadily northward from a value of about 12 mm. on the South Coast to about 2 mm. in Lancashire and Yorkshire. Over this area the average rainfall is affected little if the contribution of showers is excluded on occasions when the continuous rain area of the depression did not extend to the station concerned. Over Scotland, Northern Ireland and the extreme north of England where the average value varies irregularly between 2 and 4 mm. the main contribution to the rainfall is from showers and other minor rain areas not directly connected with the depressions which were studied. There is an absence of any significant difference between the average rainfall amounts in the east and west of Great Britain south of the Wash; further north, where the frontal rainfall of the depressions no longer dominates the map, the east coast receives more rain than the west during the passage of depressions eastward near the Channel.

In order to form a clearer picture of the rainfall in relation to the track of the depression an estimate was made of the average rainfall over areas centred 50, 150 and 250 miles to the north of the track of the centre. The resulting profiles of rainfall across the depression track are shown in Fig. 3. Western and Eastern areas of Great Britain were treated separately but there is no significant difference between their rainfall profiles. Rainfall falls off from 12 mm. at 50 miles north of the depression track to 2 mm. at 250 miles. These values are somewhat lower and the falling off steeper than for the depressions which crossed England north of the Bristol Channel (see M.R.P. 762).

Fig. 4 shows the percentage of the 42 depressions during the passage of which less than 0.1 mm. of rain fell at selected stations in the British Isles. South of the Bristol Channel and the Thames rain fell from nearly every depression. The proportion of cases without measurable rain rises to about 40% over Northern England but tends to decrease again further north as a result of the greater frequency of showers and extraneous rain not associated with the depression itself.

There is no significant difference between the number of cases without measurable rain at the high level stations and the number at adjacent stations on low ground (see inset to Fig. 4).

Fig. 5 contains histograms of rainfall amounts for selected stations and indicates the wide range of variation at individual stations among the rainfall of the various depressions which follow the selected track.

## 4. Characteristics of individual rainfall distributions

The individual rainfall charts showed a wide variety of rainfall distributions. Some of the depressions had a component of motion towards the south and affected the west of the country but missed the east; others moving towards the north gave little or no rain in western districts. However none of the depressions left the south of England entirely unaffected by the continuous rain belt.

About 2/3 of the depressions which were studied appeared to give a definite band of heavier precipitation along their tracks. However this was not easy to determine because in most cases the rainfall chart did not extend south of the track of the centre.

As with the depressions studied in M.R.P. 762 the depressions which gave a well-defined band of precipitation tended to be faster moving than the remainder, and more often associated with straight thickness lines. However, no seasonal variation appears in the proportion of depressions giving a well defined band of rainfall. All of the depressions which moved with speeds greater than 30 kts gave a definite precipitation band (about $\frac{1}{4}$ of the total) while only 18 out of the 31 depressions which moved more slowly did so. Neither this relation with speed of movement, nor the association with straight thickness lines is close enough to be of much value in forecasting.

## 5. Variation in rainfall amounts in relation to synoptic parameters

A limited study was made of the relationships between the rainfall amounts from individual depressions and parameters characteristic of the synoptic situation so chosen as to be such that a forecaster might hope to predict them by conventional methods. The similar studies of the depressions passing further to the north have given results of only very limited practical value and the present study was restricted to the simplest and most direct relationships.

In order to provide a measure of the rainfall the average value of the rainfall of six stations in Southern England was calculated for each depression; also the average for 12 stations (including the former 6) and extending as far north as Manchester. The correlation between the two averages was about + 0.9, and the average rainfall at the 12 stations was used as a rainfall parameter (R) for comparison with the synoptic situation.

It was found that depressions with a component of motion towards the north (i.e. those moving from SW to NE) tended to produce more rain than those with a component of motion towards the south. The correlation between the rainfall and the direction of motion is -0.33; a similar correlation was found in the previous investigation for depressions crossing England further north. As would be expected the rainfall over England is also greater for depressions moving across the northern part of the area in Fig. 1 than it is for those which pass further south; the correlation between the rainfall R and latitudes at which the track crosses the Greenwich meridian is + 0.47.

Investigation of the relation between rainfall and the central pressure of the depression gives a correlation coefficient of -0.35 similar to that found in the previous investigation, but little relation was found in the present data between the rainfall and the speed of the depression (correlation coefficient -0.10).

The orientation of the 1000-500 mb thickness lines have a relation with the rainfall corresponding to the relation with the track of the depression. This is brought out in Fig. 6 which shows the mean 1000-500 mb thickness charts for the 10 most rainy and the 10 least rainy depressions. The chart (6(a)) for the rainy cases shows a thermal trough to the west of the British Isles with a southerly component in the thermal wind over Southern England whereas, in contrast, the average map for the dry cases (6(b)) shows a thermal trough to the east of the country and thermal winds with a northerly component over most of the British Isles.

The partial regression equation has been calculated between the rainfall parameter R and the three parameters which gave the highest correlation among those studied. These are:-

(a) $\theta$ - orientation of the track of the depression (degs from North)
(b) $p$ - the central pressure of the depression (mb)
(c) $\ell$ - the latitude at which the track crosses the Greenwich meridian.

The regression equation is:-

$$R = 8.17 - 0.02\Theta - 0.18(p - 1000) + 1.71(\ell - 50)$$

The total correlation between R and the result of the regression equation is + 0.57. Such a correlation is not high enough to be of much practical value in forecasting, particularly when the parameters themselves have to be forecast. However, the general relationships of rainfall with the direction of motion, depth and latitude of the track of the depression are useful "background knowledge" to the forecaster.

## 6. Conclusions

Nearly all depressions moving eastward along or near to the English Channel give rain in southern England and the average rainfall amount decreases from 12 mm. on the South Coast to about 2 mm. in Lancashire and Yorkshire. Further north although some rain usually falls during the passage of a depression up the Channel and averages 2 to 3 mm. it is not directly connected with the depression.

Orographic influences do not have any very pronounced effect on the rainfall of these depressions and there is little systematic difference in rainfall amounts between the east and west of Great Britain.

A well defined band of heavier rainfall is found along the track of more than half of the depressions. It is most clearly defined with depressions moving at more than 30 kts and with a "straight" thickness pattern. The present study does not confirm the seasonal variation in the incidence of the precipitation band which was suggested in an earlier investigation.

The investigation confirms the tendency already recognised for the rainfall to be greater with the deeper depressions and with those which have a component of motion towards the north. The negative correlation between rainfall and the speed of the depression which was found in previous investigations is not well developed in the present data. The factor which has the closest association with the rainfall is the latitude of the track of the centre, measured here at its intersection with the Greenwich meridian. This relation is simply an expression of the fact that the nearer the centre passes to Southern England the greater the rainfall there, but it emphasises that the forecasting of rainfall amounts from prebaratic charts depends on an accurate pre-knowledge of the detailed behaviour of depression centres.

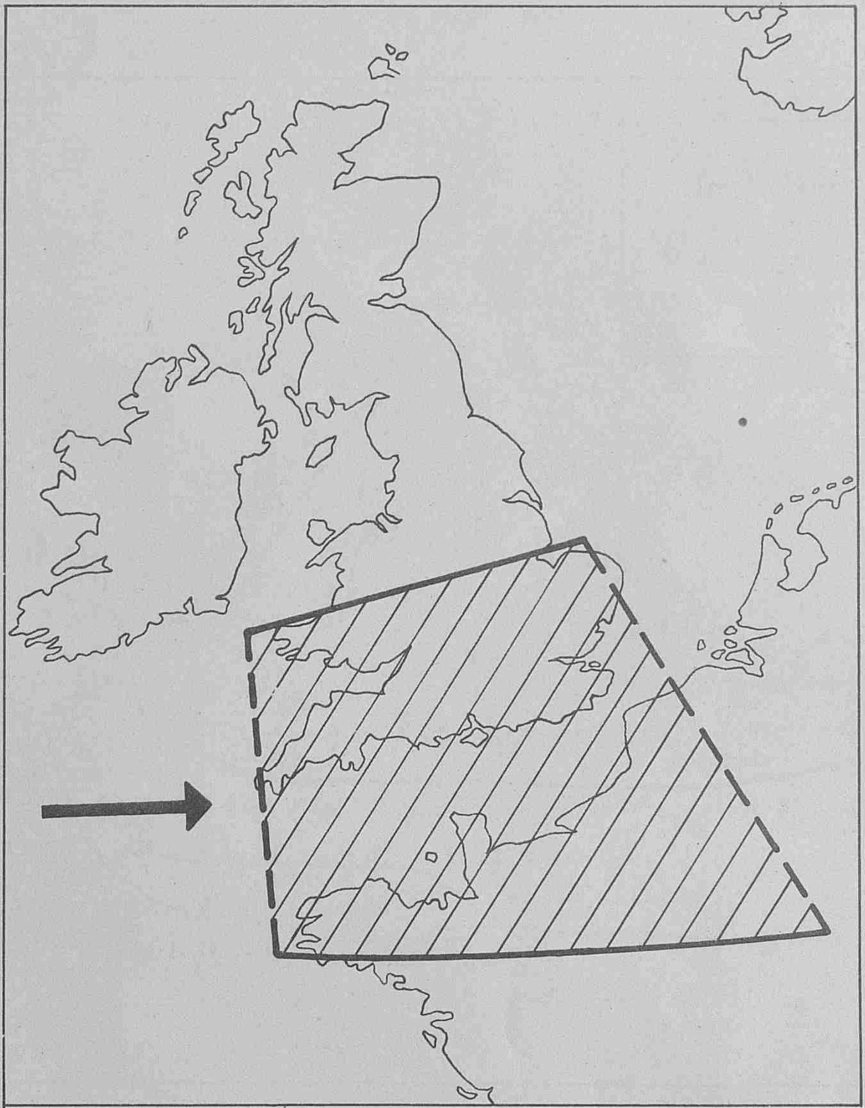


Fig. 1 Area through which tracks of depressions were required to pass

M.R.P. 816.

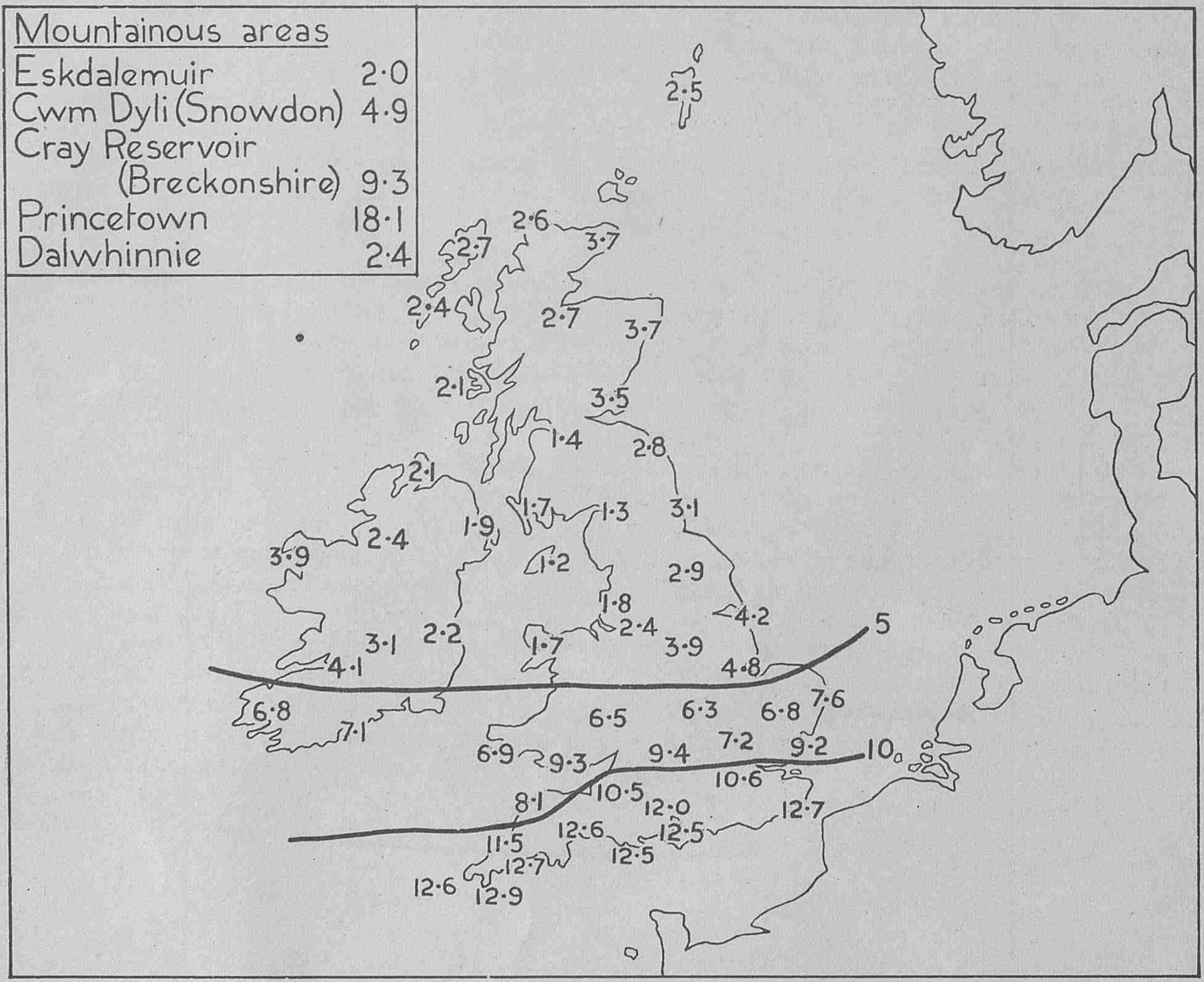

Mountainous areas
Eskdalemuir 2·0
Cwm Dyli (Snowdon) 4·9
Cray Reservoir (Breckonshire) 9·3
Princetown 18·1
Dalwhinnie 2·4
2·5
2·6
3·7
2·7
2·4
2·7
3·7
2·1
3·5
1·4
2·8
2·1
1·9
1·7
1·3
3·1
2·4
3·9
1·2
2·9
1·8
4·2
2·4
5
3·1
2·2
1·7
3·9
4·1
4·8
7·6
6·8
6·5
6·3
6·8
7·1
7·2
9·2
6·9
9·3
9·4
10
10·6
10·5
8·1
12·0
12·7
12·6
12·5
11·5
12·7
12·5
12·6
12·9

Fig. 2 Average rainfall (m m.) from depressions which moved eastward near to the English Channel — 42 cases — period 1941–1950

M.R.P. 816.

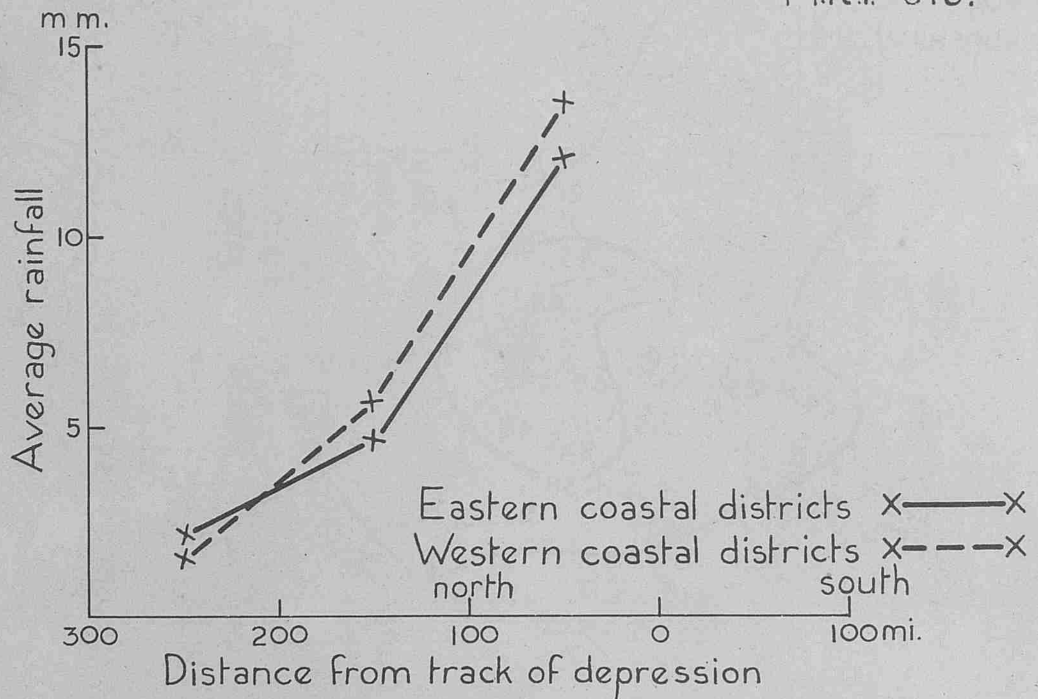
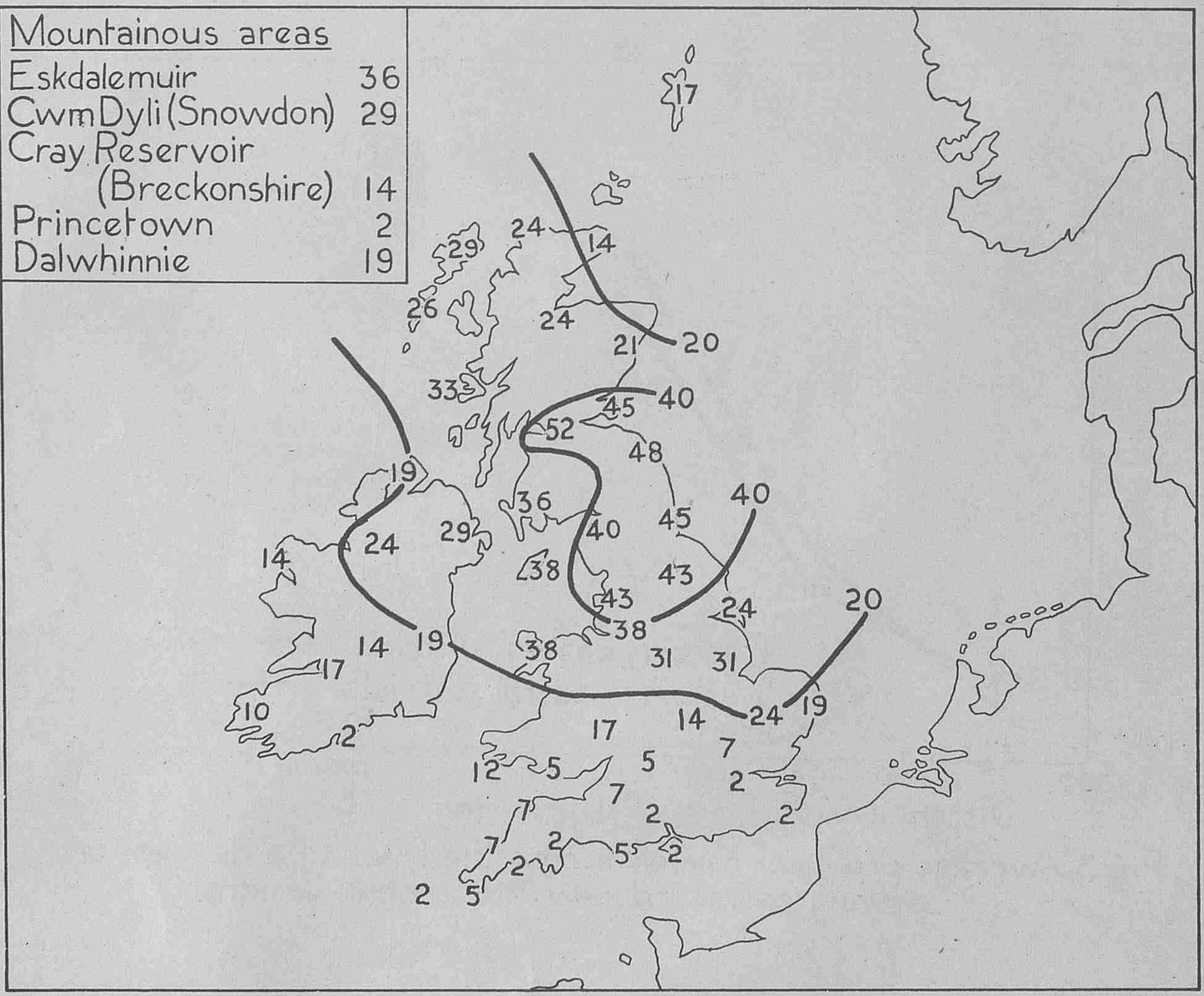


Fig. 3 Average profile of rainfall across the track of a depression moving eastward near the English Channel



| Mountainous areas | |
|---|---|
| Eskdalemuir | 36 |
| Cwm Dyli (Snowdon) | 29 |
| Cray Reservoir (Breckonshire) | 14 |
| Princetown | 2 |
| Dalwhinnie | 19 |

17
24
14
29
26
24
21
20
33
45
40
52
48
19
36
40
45
40
24
29
14
38
43
43
24
20
38
14
19
38
31
31
17
10
2
17
14
24
19
7
12
5
5
7
2
2
7
2
7
5
2
2
2
5
2

Fig. 4 Percentage of depressions moving eastward near the English Channel which gave less than 0·1 mm. of rain — 42 cases — period 1941-1950

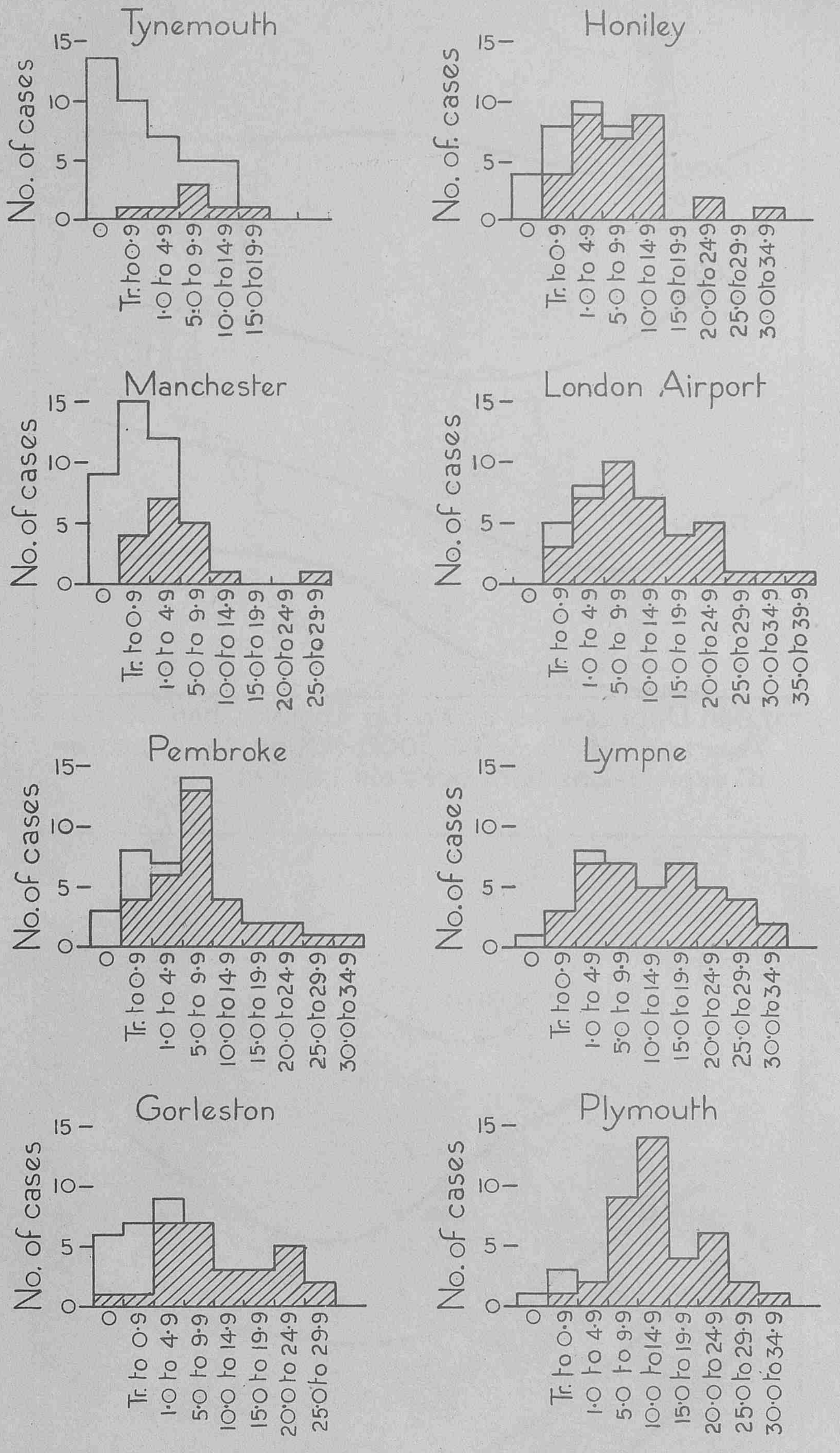


Fig. 5 Depressions which moved eastward near the English Channel — 42 cases — period 1941-1950
Frequency diagrams of rainfall totals (m.m.) from individual depressions at selected stations. Shaded area indicates frequency based only on occasions when continuous rain area affected station

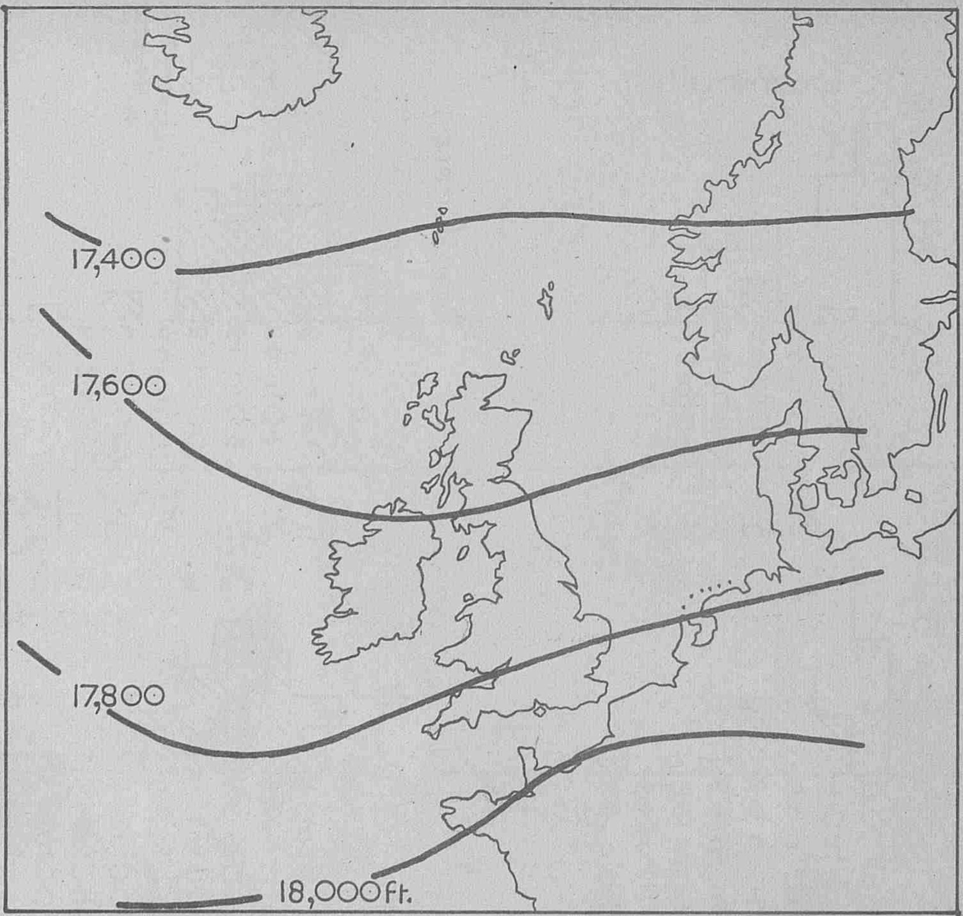

17,400
17,600
17,800
18,000 ft.

Fig. 6(a) Depressions going up English Channel — Average distribution of 1000-500 mb. thickness for 10 depressions with greatest rainfall

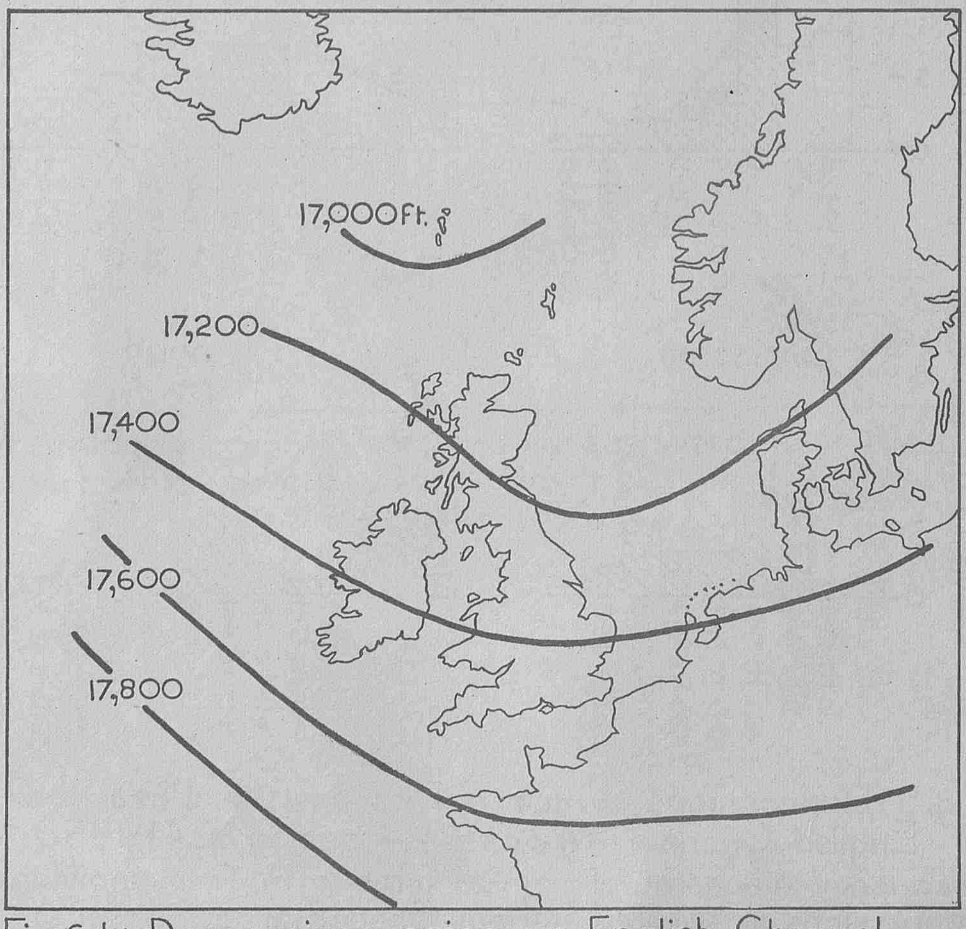

17,000 ft.
17,200
17,400
17,600
17,800

Fig. 6(b) Depressions going up English Channel — Average distribution of 1000-500 mb. thickness for 10 depressions with least rainfall

MRP's 817, 818, 819, 860, 820

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M.R.P. 817

S.C.III/154

29th May, 1953.

AD 44246

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# An interim report on the measurement of the vertical distribution of atmospheric ozone by a chemical method, to heights of 12 km., from aircraft

by

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Clarendon Laboratory,
Oxford.

### Introduction.

These measurements have five main objects:-

(1) to facilitate, by providing reliable ozone concentration values for the first 12 km. of the atmosphere, the difficult computation of the vertical distribution of ozone through the whole atmosphere as determined by spectrophotometric measurements of zenith scattered skylight in the ultra-violet; the so-called Gotz "umkehr" method.

(2) to investigate how ozone concentration varies with height as the tropopause is crossed. This observation, together with simultaneous frost point measurements at the same levels, might help to fix orders of magnitude for the rate and vertical extent of any regular exchange of air there might be between stratosphere and troposphere. One may also be able to test the probability of any proposed long period general circulation extending to the ozone-rich regions of the atmosphere, attempting to use ozone as a "tracer" gas.

(3) to investigate how the ozone concentration up to 12 km. varies with season and weather conditions and to compare this with the now well known total ozone changes. Total ozone values are available from the Dobson Spectrophotometer network in England.

(4) to stimulate interest in reliable and comparable measurements of low level ozone amounts in other parts of the world, particularly in high and low latitudes and preferably by the same method. This would determine the latitude variation of low level ozone more certainly than do the existing diversity of methods used.

(5) to show the need for an extension of this work to greater heights.

### Choice of method.

For the measurement of ozone concentrations locally at particular levels in the atmosphere, spectrophotometric methods, though they can be made specific to ozone, become very insensitive. They require, because of the very low concentration of the gas, optical path lengths of the order of 1 km. Furthermore such techniques demand the greatest of care in instrumentation even at ground level. It is difficult to envisage a reliable airborne installation.

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One is obliged to turn to chemical methods which are not in general, specific to ozone. However there does exist one well tried and fully documented reaction - the oxidation of a heavily buffered neutral aqueous solution of potassium iodide. This technique, when used with certain precautions fully described by Ehmert, becomes, though not entirely specific to ozone, by a very large factor more efficient in the detection of that gas than of the oxides of nitrogen and other oxidising constituents of the atmosphere.

An apparatus, designed around Ehmert's method has been constructed at Oxford jointly by Prof. Dobson, Dr. Brewer and the writer, and has been used in Mosquito aircraft of the Met. Research Flight at Farnborough. This measures ozone concentration at 15 or 16 selected levels of the atmosphere up to the ceiling of the aircraft. Two or three reaction vessels are kept as controls and not exposed to the air stream.

## Method.

The basic reaction is the oxidation of potassium iodide, liberating free iodine:-

$$O_3 + 2KI + H_2O = 2KOH + O_2 + I_2$$

Small amounts of sodium thiosulphate solution are added before sampling the ozone. This reagent, immediately absorbing any free iodine, prevents any loss of that substance by evaporation.

$$2Na_2S_2O_3 + I_2 = 2NaI + Na_2S_4O_6$$

One determines the amount of free iodine liberated by the ozone in a controlled volume of air by electrometrically titrating the amount of thiosulphate remaining unconverted to tetrathionate.

Five litres of air are sampled at each level. The air flow through the sampling vessel is controlled by a relief valve keeping constant the pressure drop across a capillary tube in series with the sampling vessel, over a wide range of input air pressure. The operator has merely to expose each sampling vessel in turn to the air stream for a fixed time (5 min.).

## Tests of the method.

(1) Two similar but portable air-samplers were taken in March to Prof. Dobson's observatory at his farm, "High Leys" near Kendal in Westmorland. The ozone concentration was measured chemically at two stations, one at "High Leys" and the other at an adjacent farm "High House" about 1 km. away.

At the same time, the attenuation of the 2536AU mercury line which is strongly absorbed by ozone, was measured along this 1 km. optical path, relative to the attenuation of the weakly absorbed line 2967AU.

The light source was an under run, medium pressure mercury lamp at the focus of a 6" focal length, 16" diameter, front aluminised paraboloid. The spectrographs were the Dobson photographic Fery type instruments, used originally for the first world wide ozone survey. The Lake District was chosen for its clear atmosphere. Two spectrographs were used, one to measure the intensity differences at each end of the path, the other as a control to monitor any change in the light source itself.

The ozone amounts measured chemically ranged between c.0.001 and 0.003 cm./km. These values, plotted against the measured optical attenuation over the 1 km. path, fall, within the experimental error, very closely upon a straight line of the expected slope. (Fig. 1). The elimination of all systematic error in the base line of the spectrographically determined attenuation, would require a somewhat lengthy programme of work. (For example, it is not an easy matter to make the final images in the spectrograph identical in focus and completely uniform at two stations separated by a long optical path.) Nevertheless, the differences of attenuation for differing ozone amounts are quite well determined and the reasonable fit of the observations to a line of the expected slope, over the good range of ozone amount encountered, gives rise to confidence in the chemical method used here.

(2) The constant air flow device has been tested only on the ground. The flow remains constant through the capillary to within ± 5% as the input air pressure varies between c.6 and c.100 cm. $H_2O$.

The constant pressure relief valve maintains the same pressure differences with hydrogen flowing as with air and is evidently largely independent of gas density. There is thus every reason for confidence in the control of the air flow. Before the experiments are completed it is planned to test this in flight unambiguously (collecting the air over water in a measuring cylinder.)

Results.

There have been 13 successful flights since October 1952. These are continuing and it is proposed to present first results here with the minimum of comment - awaiting a further occasion for a more detailed report.

Figs. 2 - 8a show, as a function of height, the ozone amount expressed in two different ways (1) as a concentration in cm./km. (2) as a mixing ratio, cc of $O_3$/cc of air. In addition are plotted the temperatures and frost points (dashed lines) measured at the same time, and the cloud amounts observed. (In some cases the temperature data are incomplete, these ascents are supplemented by the appropriate radiosonde data (dotted lines)).

The probable error of the ozone measurements is indicated by the cross bars - a smooth curve is then drawn through these points and this curve is converted into the mixing ratio curve.

Individual days sometimes show considerable structure in the ozone concentration profile. This will be worthy of further comment when more flights are completed.

To obtain first some orders of magnitude, all the ozone profiles were meaned at 1,000 ft. intervals and are plotted in Fig. 8b. An estimate of the representativeness of this mean curve is given by the rather small scatter of the individual mean points. (The agreement with three ascents by Ehmert, using the same reaction but a different apparatus, Fig. 8b, underlines the reliability of the method.)

These mean curves show several interesting features:

(1) The amount of ozone below 12 km. is on the average about 8% of the total ozone. (See also Table I for the scatter of individual values.)

(2) There appears to be no regular discontinuity in the ozone mixing ratio at the tropopause.

These two observations are already of value, for the Gotz "umkehr" method is rather insensitive in fixing the amount of low ozone. These observations fix the amount and mean vertical distribution for the first 12 km. of the atmosphere in our latitude, leaving one less unknown in the already overburdened computations necessary to arrive at a vertical distribution from zenith skylight measurements. The mean curve is again plotted in Fig. 9 together with some representative vertical distributions by the "umkehr" method and balloon spectrograph. The fit with the "umkehr" measurements is satisfactory but more weight should be given to the chemical determinations at these levels than to the spectrophotometric. Result (2) which certainly could not be established by zenith skylight measurements, is of interest in discussing possible general circulation patterns in the atmosphere. Water vapour measurements with the frost point hygrometer invariably show the lower stratosphere to be very dry in England. This seems to demand a general circulation in which the air in the lower stratosphere in our latitudes has previously passed through regions near the equatorial tropopause. It can there be dried sufficiently by the very low temperatures and precipitation.

The circulation postulated to bring this air to the lower stratosphere of our latitudes may or may not imply a trajectory sufficiently high for the air to have entered ozone-rich regions of the atmosphere (hts. between 20 & 30 km.). In the former case one might expect to find the air of the lower stratosphere somewhat enriched in ozone content, just as it is very much drier than the air below the tropopause. That this is not markedly so (see Fig. 10 the mean vertical distribution from these ascents, now computed at heights relative to the tropopause and not ground level) may favour such a circulation's being limited in the height to which it may extend to levels lower than about 20-25 km.

Before one can be definite on this matter more flights are necessary. Thirteen flights can not be expected to be comprehensive nor fully representative of even average conditions. It will be noticed that in these ascents there has not yet been observed the very sharp discontinuity in frost point at the tropopause which is sometimes detected. Nor is the generally observed increase in lapse rate of the frost point in the stratosphere apparent in the mean curve (Fig. 10). One is also limited to days of tropopause lower than the ceiling of the aircraft. A close examination of the assumptions necessary in using ozone as a tracer in this connection is essential before more than tentative conclusions are published.

## Remarks.

Sufficient results are already available to show the interest in this type of measurement. A glance at Fig. 9 is sufficient to show the value of extending the measurements to greater heights. It is hoped to instal the equipment as soon as possible in the Canberra aircraft which is coming into service for the M.R.F. If the "umkehr" measurements are correct at these levels, we should then be able to detect the beginning of the rise in ozone concentration towards its maximum at about 22 km.

The next development, after the first "Canberra" results, ought to be the development of an automatically recording instrument. This should not be too difficult and would allow the measurements to become routine. A great deal of care is necessary with the present system of sampling and analysis and it is not particularly suited to rapid routine work. At a later stage it may be possible to develop a balloon borne equipment. This step is a major one and though probably it would result in a rather heavy equipment requiring many or even specially made balloons, it may be thought a worth while undertaking because of the greater heights attainable. It is unlikely that the cost of even an elaborate balloon borne equipment would much exceed the cost of an aircraft ascent with its lower attainable "ceiling."

## Acknowledgements.

To the Gassiot Committee of the Royal Society for a research grant. To the Met. Research Flight, in particular Mr. R. Murgatroyd for his ready

help and interest, Messrs. P. Goldsmith and W. Hollings who operated the equipment and Fg.Off. A. Bavin who was pilot for most of these flights. To the Commanding Officers of Benson and Abingdon R.A.F. stations for allowing the use of these airfields.

Professor Dobson and Dr. A.W. Brewer gave much attention to the detailed design of the airborne apparatus, which was constructed in the Clarendon Laboratory. Thanks are due to Professor Lord Cherwell for making these laboratory services available and to Dr. A.W. Brewer for his considerable help with transport to both "High Leys" and Farnborough.

## References.

The recent background to the topics discussed here is summarised in the proceedings of the Oxford symposium on atmospheric ozone published in the U.G.G.I. News Letter of April 1953.

Earlier work and full bibliographies are available in the American Compendium of Meteorology, particularly Gotz on Atmospheric Ozone and Dobson and Brewer on Water Vapour in the Lower Stratosphere.

### Table I

### A summary of the total amounts of low level ozone measured in the individual ascents.

| DATE 1952 | TOTAL OZONE x. cm. | OZONE AMOUNT below y. cm. | Km. | y/x% | AVERAGE CONC. cm./km. |
|---|---|---|---|---|---|
| Oct.21 | .227 | .024 | 9.5 | 10½ | .002(5) |
| Nov.12 | .190 | .024 | 11.7 | 12½ | .002 |
| Nov.20 | .207 | .008 | 11.5 | 4 | .001 |
| Nov.24 | .231 | .018 | 11.5 | 7½ | .001(5) |
| 1953 | | | | | |
| Jan.30 | .224 | .030 | 12.0 | 13½ | .002(5) |
| Feb.2 | .232 | .010 | 12.0 | 4½ | .001 |
| Feb.4 | .260 | .021 | 11.5 | 8 | .002 |
| Feb.13 | .294 | .018 | 12.5 | 6 | .001(5) |
| Feb.16 | .270 | .018 | 12.2 | 6½ | .001(5) |
| Feb.18 | .234 | .013 | 12.0 | 5½ | .001 |
| Mar.11 | .280 | .024 | 12.0 | 8½ | .002 |
| Mar.13 | .274 | .022 | 12.2 | 8 | .002 |
| May.11 | .295 | .022 | 12.0 | 7½ | .002 |
| MEANS:- | .247 | .019 | 11.7 | 8 | .001(7) |

Remarks. The three days of lowest average concentration, Nov. 20th, 1952, Feb. 2nd and Feb. 18th, 1953, are all characterised by remarkable deficiencies in ozone in the levels nearest the ground. This is possibly due to highly polluted air or stagnant air at those levels. These are also the three days when the measured ozone amount is the lowest fraction of the total ozone in the atmosphere.

November 12th and January 30th, characterised by high measured amounts relative to the total ozone x, are each associated with large maxima in the troposphere, in each case possibly associated with stratified cloud:- an effect also noticed and attributed perhaps to electrical phenomena by Ehmert. Not infrequently Cirrus cloud seems to be associated with a rather marked minimum in the ozone concentration profile. Again more ascents are needed to be definite. On the whole there is little correlation between low level ozone amounts and total ozone, for these few ascents. The amount of ozone below about 12 km. is around 8% of the total, on the average.

M.R.P. 817

Spectrographic
Test of the Method

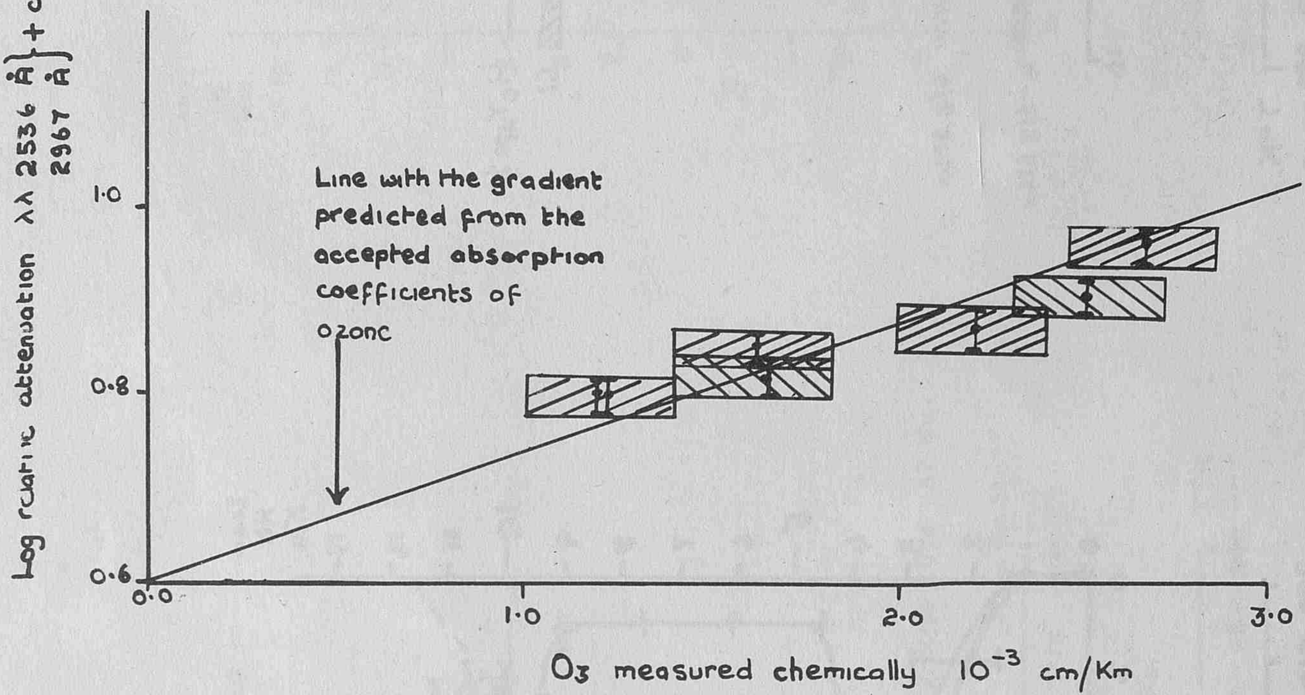
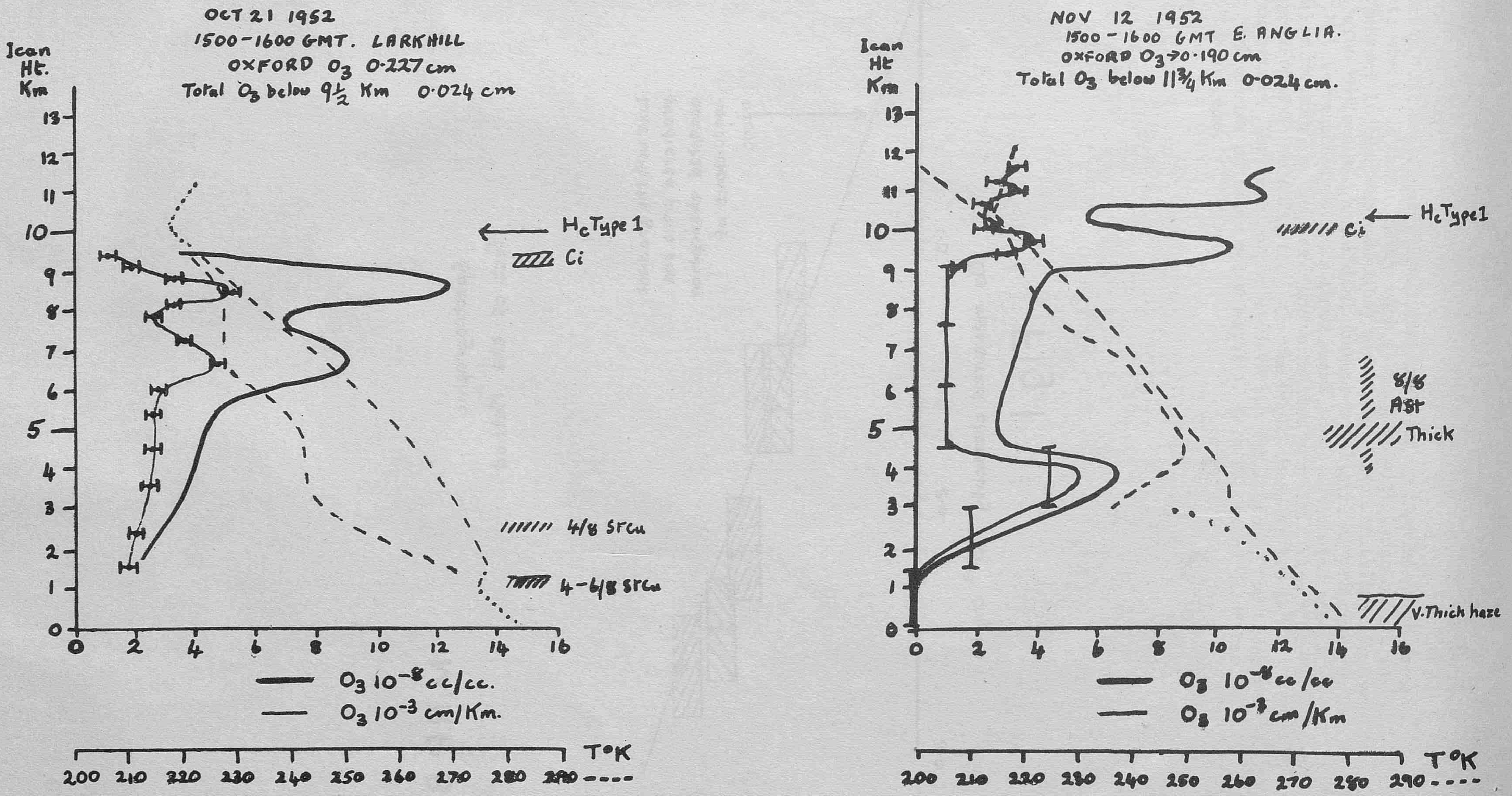


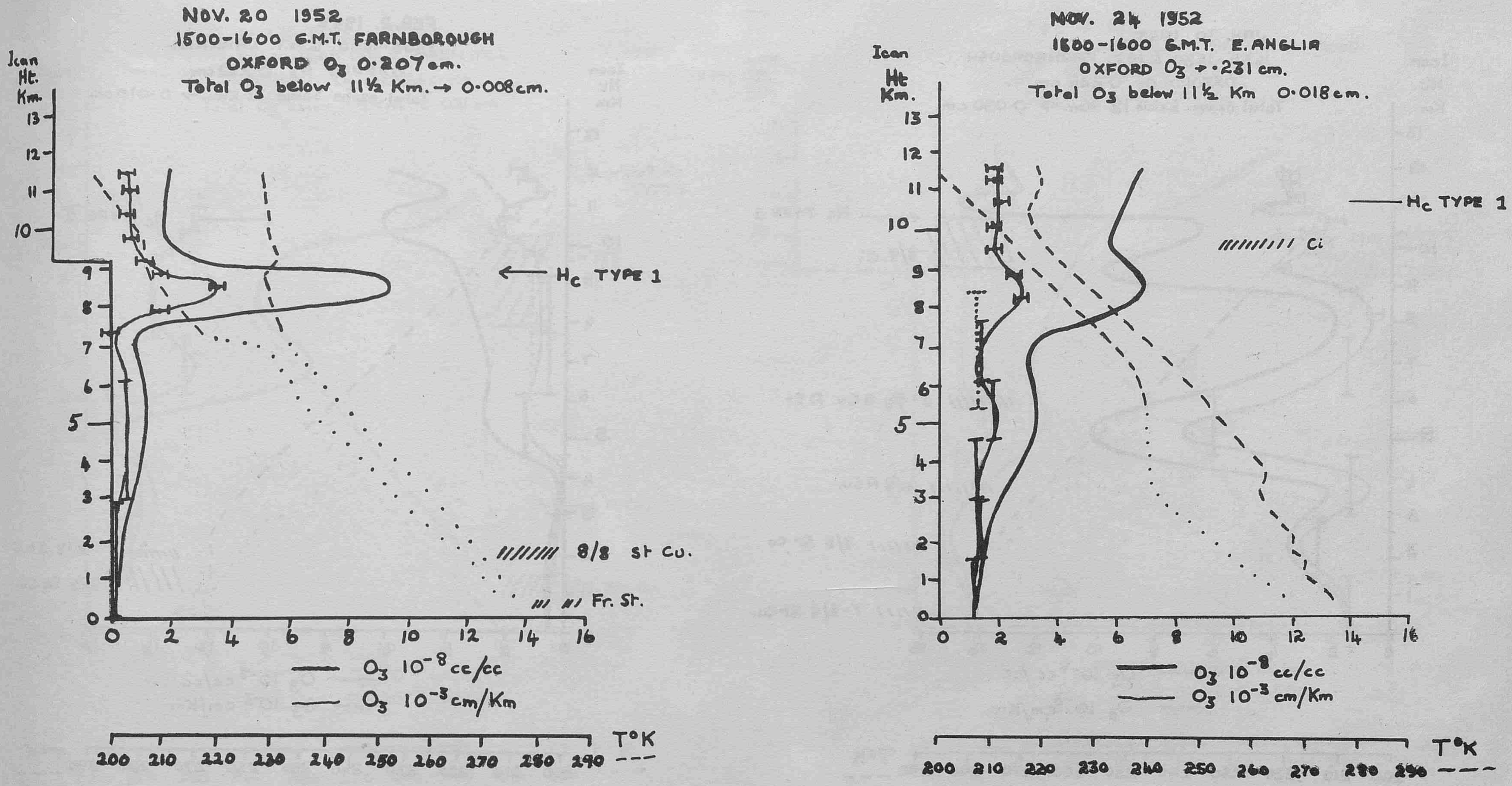
Fig. 1


OCT 21 1952
1500-1600 GMT. LARKHILL
OXFORD $O_3$ 0·227 cm
Total $O_3$ below 9½ Km 0·024 cm
Ican Ht. Km
13
12
11
10
9
8
7
6
5
4
3
2
1
0
0
2
4
6
8
10
12
14
16
$H_c$ Type 1
Ci
4/8 Srcu
4-6/8 Srcu
$O_3$ $10^{-8}$ cc/cc.
$O_3$ $10^{-3}$ cm/Km.
T°K
200
210
220
230
240
250
260
270
280
290
NOV 12 1952
1500-1600 GMT E. ANGLIA.
OXFORD $O_3$ > 0·190 cm
Total $O_3$ below 11¾ Km 0·024 cm.
Ican Ht Km
$H_c$ Type 1
Ci
8/8 Ast
Thick
V. Thick haze
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
T°K

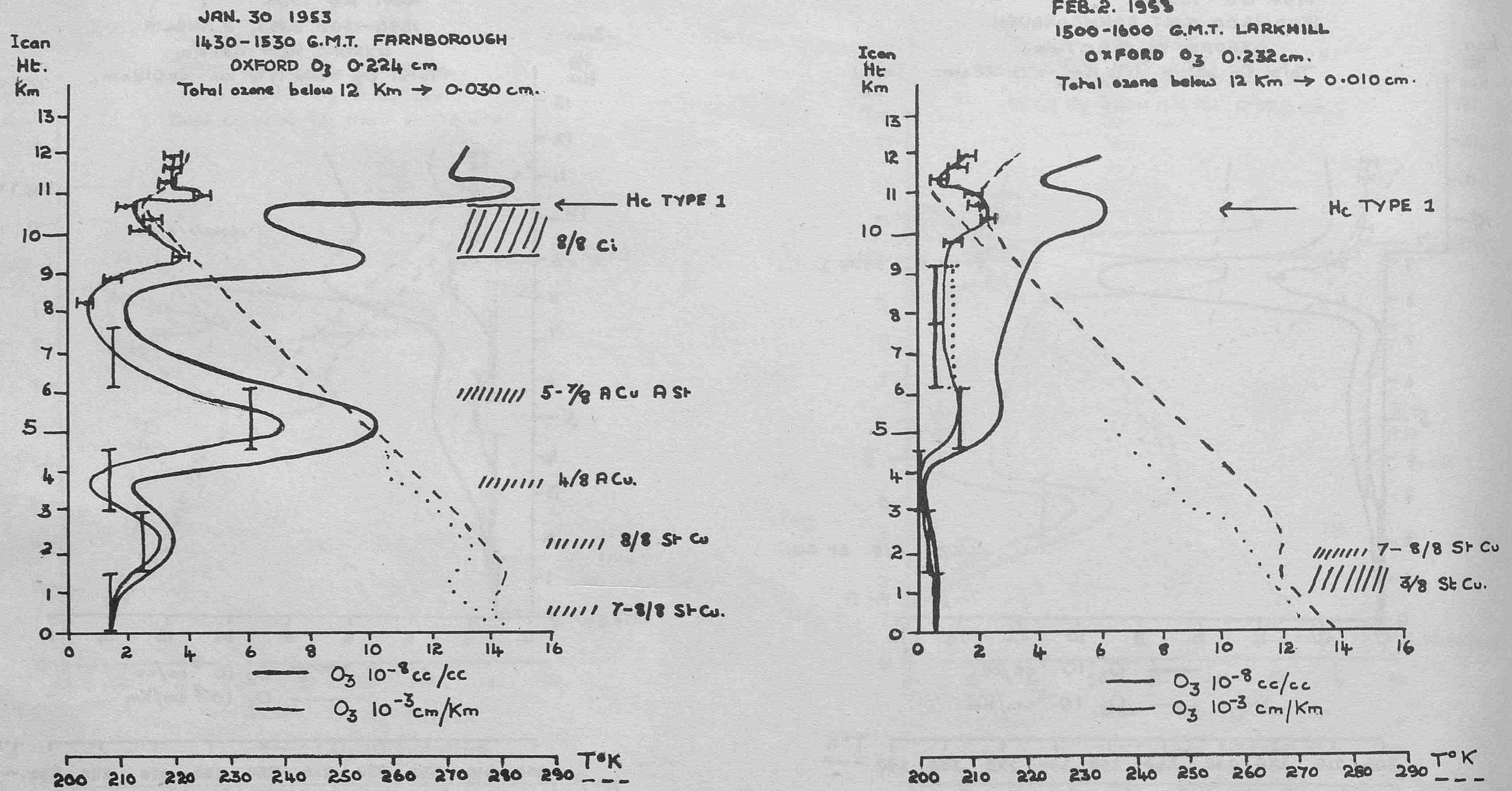
(a)

(b)

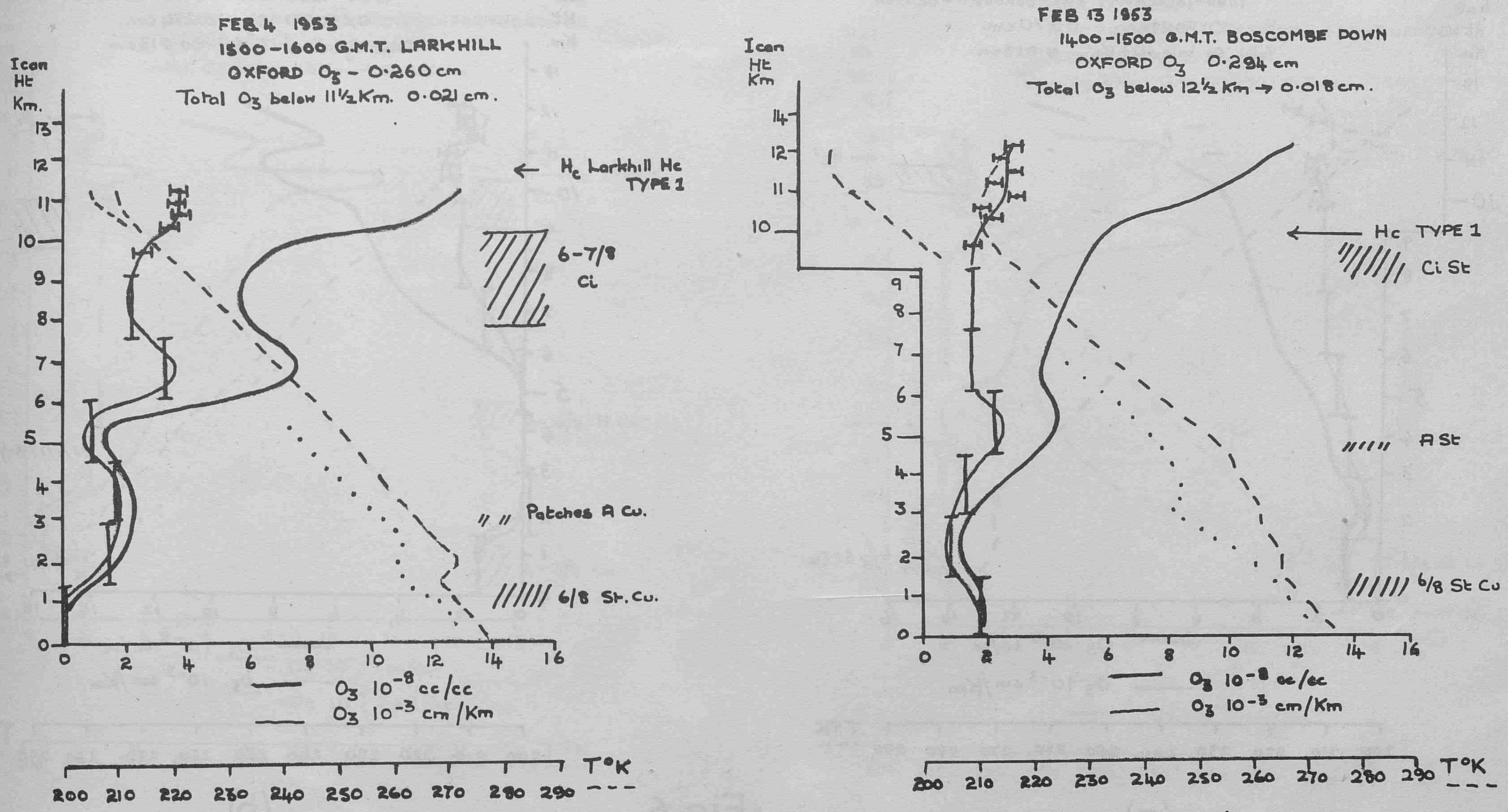
Fig. 2


NOV. 20 1952
1500-1600 G.M.T. FARNBOROUGH
OXFORD $O_3$ 0·207 cm.
Total $O_3$ below 11½ Km. → 0·008 cm.
Ican Ht. Km.
13
12
11
10
9
8
7
6
5
4
3
2
1
0
0
2
4
6
8
10
12
14
16
← $H_c$ TYPE 1
8/8 St Cu.
Fr. St.
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
T°K
200
210
220
230
240
250
260
270
280
290
NOV. 24 1952
1500-1600 G.M.T. E. ANGLIA
OXFORD $O_3$ → ·231 cm.
Total $O_3$ below 11½ Km 0·018 cm.
Ican Ht Km.
$H_c$ TYPE 1
Ci
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
T°K

(a) Fig. 3 (b)


JAN. 30 1953
1430-1530 G.M.T. FARNBOROUGH
OXFORD $O_3$ 0·224 cm
Total ozone below 12 Km → 0·030 cm.
Ican Ht. Km
13
12
11
10
9
8
7
6
5
4
3
2
1
0
0
2
4
6
8
10
12
14
16
Hc TYPE 1
8/8 Ci
5-7/8 ACu ASt
4/8 ACu.
8/8 St Cu
7-8/8 St Cu.
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
200
210
220
230
240
250
260
270
280
290
T°K
FEB. 2. 1953
1500-1600 G.M.T. LARKHILL
OXFORD $O_3$ 0·232 cm.
Total ozone below 12 Km → 0·010 cm.
Ican Ht Km
Hc TYPE 1
7-8/8 St Cu
3/8 St Cu.
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
T°K

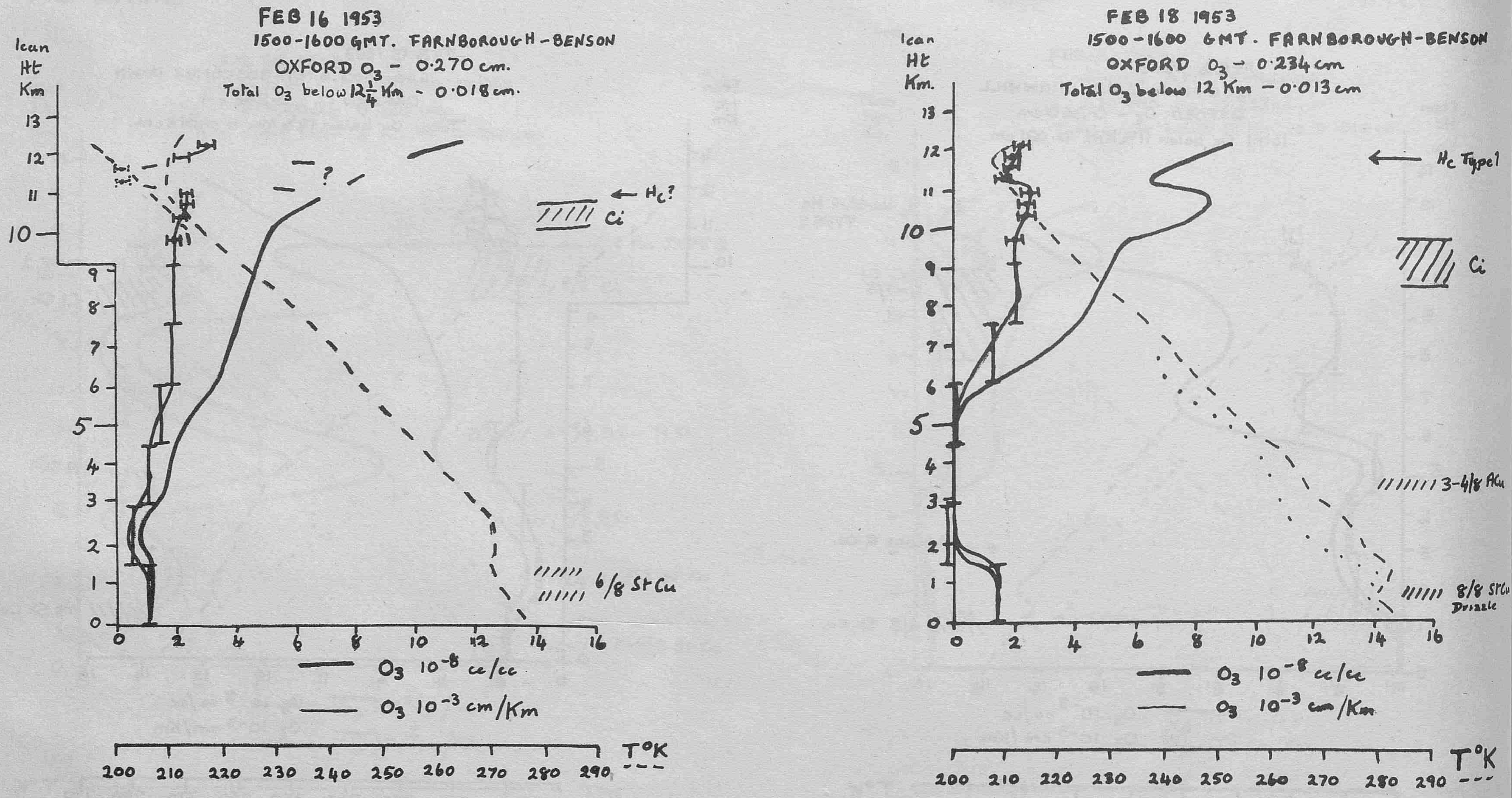
(a) Fig. 4 (b)


FEB.4 1953
1500-1600 G.M.T. LARKHILL
OXFORD O3 - 0·260 cm
Total O3 below 11½ Km. 0·021 cm.
Ican Ht Km.
0
1
2
3
4
5
6
7
8
9
10
11
12
13
0 2 4 6 8 10 12 14 16
← Hc Larkhill Hc TYPE 1
6-7/8 Ci
Patches A Cu.
6/8 St. Cu.
O3 10⁻⁸ cc/cc
O3 10⁻³ cm/Km
T°K
200 210 220 230 240 250 260 270 280 290
FEB 13 1953
1400-1500 G.M.T. BOSCOMBE DOWN
OXFORD O3 0·294 cm
Total O3 below 12½ Km → 0·018 cm.
Ican Ht Km
0
1
2
3
4
5
6
7
8
9
10
11
12
14
0 2 4 6 8 10 12 14 16
← Hc TYPE 1
Ci St
A St
6/8 St Cu
O3 10⁻⁸ cc/cc
O3 10⁻³ cm/Km
T°K
200 210 220 230 240 250 260 270 280 290

(a)

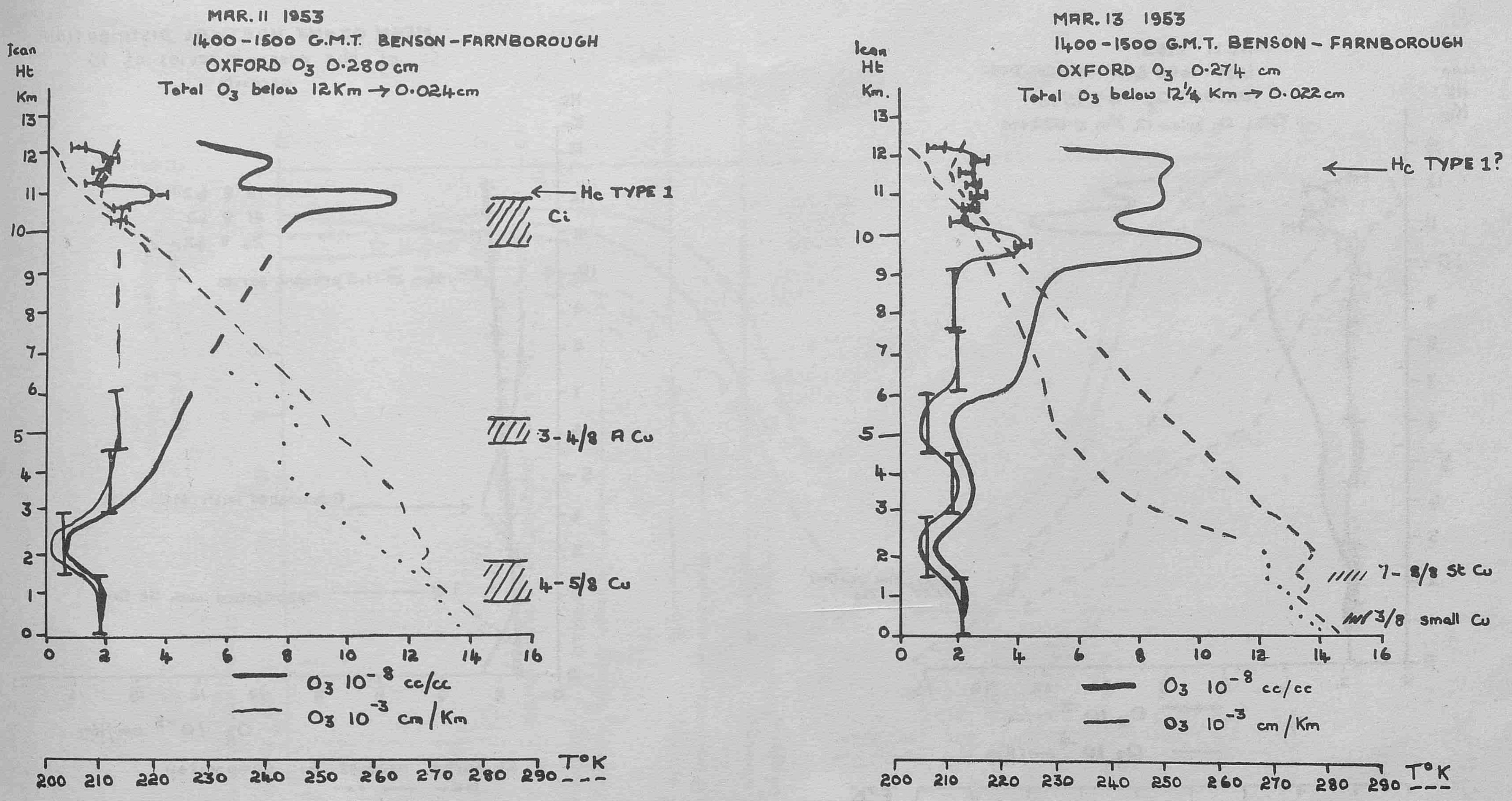
(b)

Fig.5


FEB 16 1953
1500-1600 GMT. FARNBOROUGH-BENSON
OXFORD $O_3$ - 0·270 cm.
Total $O_3$ below $12\frac{1}{4}$ Km - 0·018 cm.
Ican Ht Km
13
12
11
10
9
8
7
6
5
4
3
2
1
0
?
← $H_c$?
Ci
6/8 St Cu
0
2
4
6
8
10
12
14
16
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
T°K ---
200
210
220
230
240
250
260
270
280
290
FEB 18 1953
1500-1600 GMT. FARNBOROUGH-BENSON
OXFORD $O_3$ - 0·234 cm
Total $O_3$ below 12 Km - 0·013 cm
Ican Ht Km.
13
12
11
10
9
8
7
6
5
4
3
2
1
0
← $H_c$ Type 1
Ci
3-4/8 Acu
8/8 StCu Drizzle
0
2
4
6
8
10
12
14
16
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
T°K ---
200
210
220
230
240
250
260
270
280
290

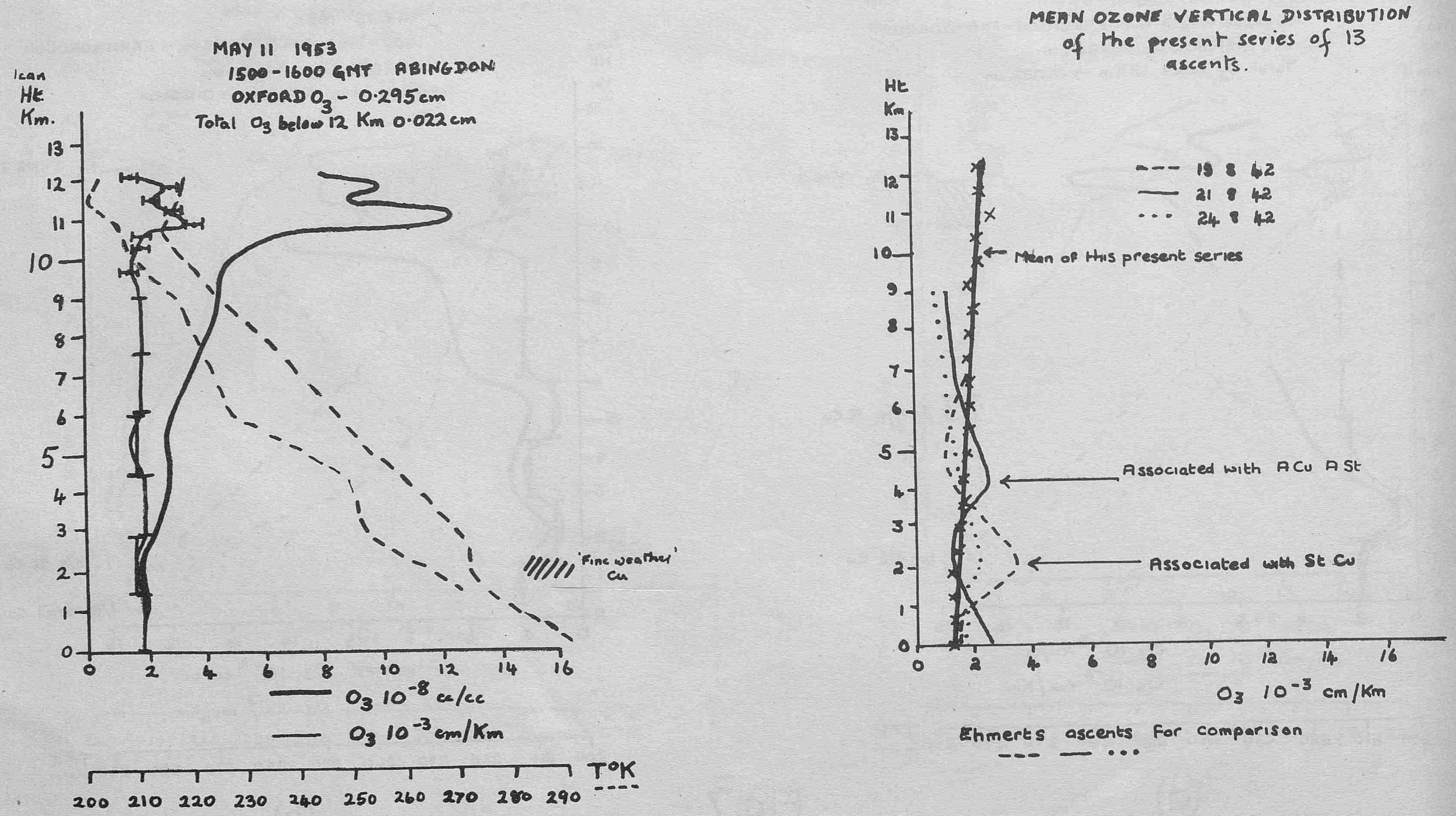
(a) (b)

Fig. 6


MAR. 11 1953
1400-1500 G.M.T. BENSON-FARNBOROUGH
OXFORD O3 0·280 cm
Total O3 below 12 Km → 0·024 cm
Ican Ht Km
0 1 2 3 4 5 6 7 8 9 10 11 12 13
0 2 4 6 8 10 12 14 16
Hc TYPE 1
Ci
3-4/8 A Cu
4-5/8 Cu
O3 10^-8 cc/cc
O3 10^-3 cm/Km
200 210 220 230 240 250 260 270 280 290
T°K
MAR. 13 1953
1400-1500 G.M.T. BENSON - FARNBOROUGH
OXFORD O3 0·274 cm
Total O3 below 12¼ Km → 0·022 cm
Ican Ht Km.
Hc TYPE 1?
7-8/8 St Cu
3/8 small Cu

(a) (b)

Fig. 7


MAY 11 1953
1500-1600 GMT ABINGDON
OXFORD $O_3$ - 0·295 cm
Total $O_3$ below 12 Km 0·022 cm
Ican Ht Km.
13
12
11
10
9
8
7
6
5
4
3
2
1
0
0
2
4
6
8
10
12
14
16
$O_3$ $10^{-8}$ cc/cc
$O_3$ $10^{-3}$ cm/Km
200
210
220
230
240
250
260
270
280
290
T°K
'Fine weather' Cu
MEAN OZONE VERTICAL DISTRIBUTION of the present series of 13 ascents.
Ht Km
19 8 42
21 8 42
24 8 42
Mean of this present series
Associated with ACu ASt
Associated with St Cu
$O_3$ $10^{-3}$ cm/Km
Ehmerts ascents for comparison

(a)

(b)

Fig. 8

Some typical $O_3$ vertical distribution profiles

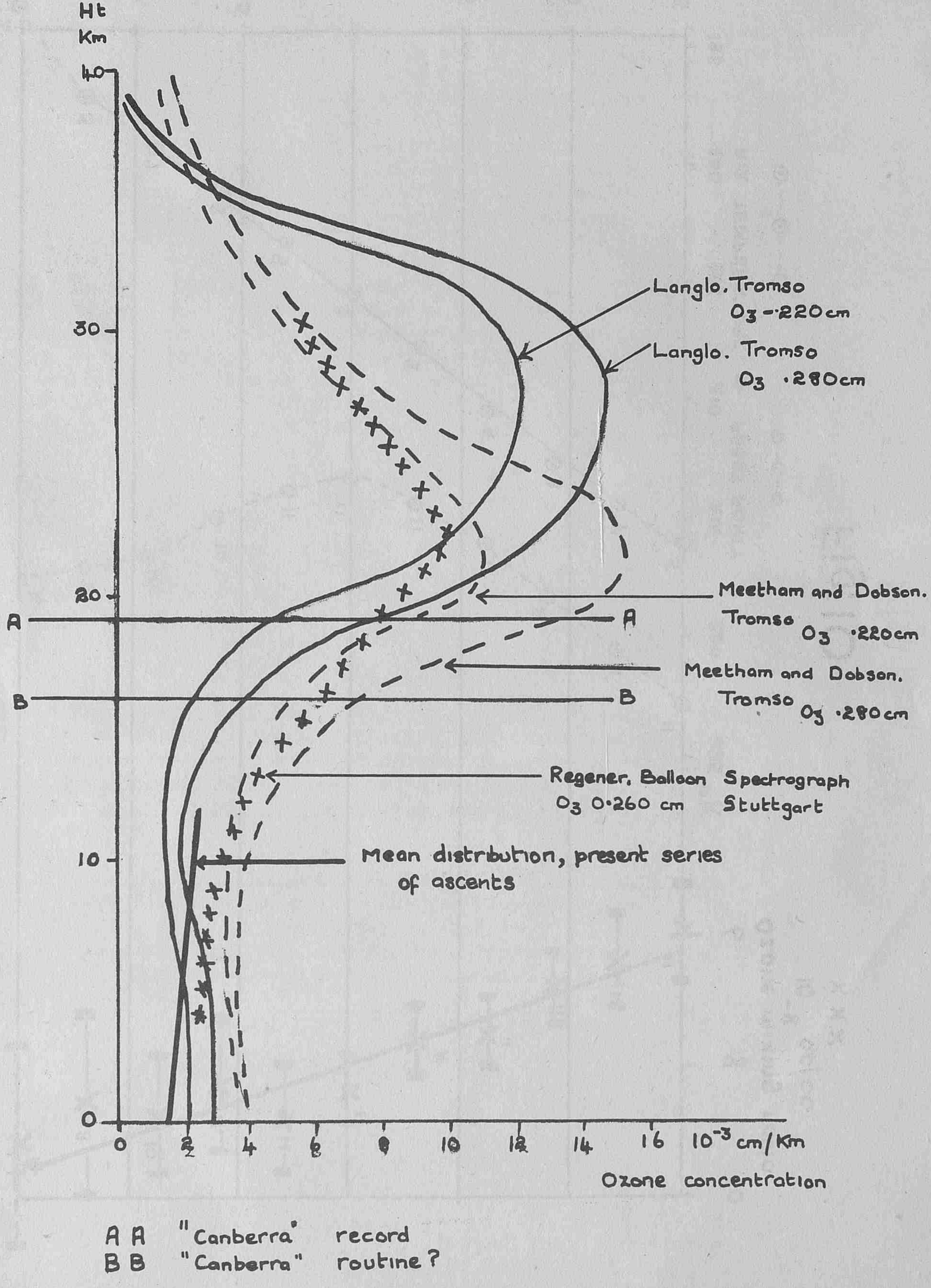


Fig. 9

MEAN VALUES of frost point, air temperature and ozone concentration relative to the tropopause

(No. of observations indicated beside each point)

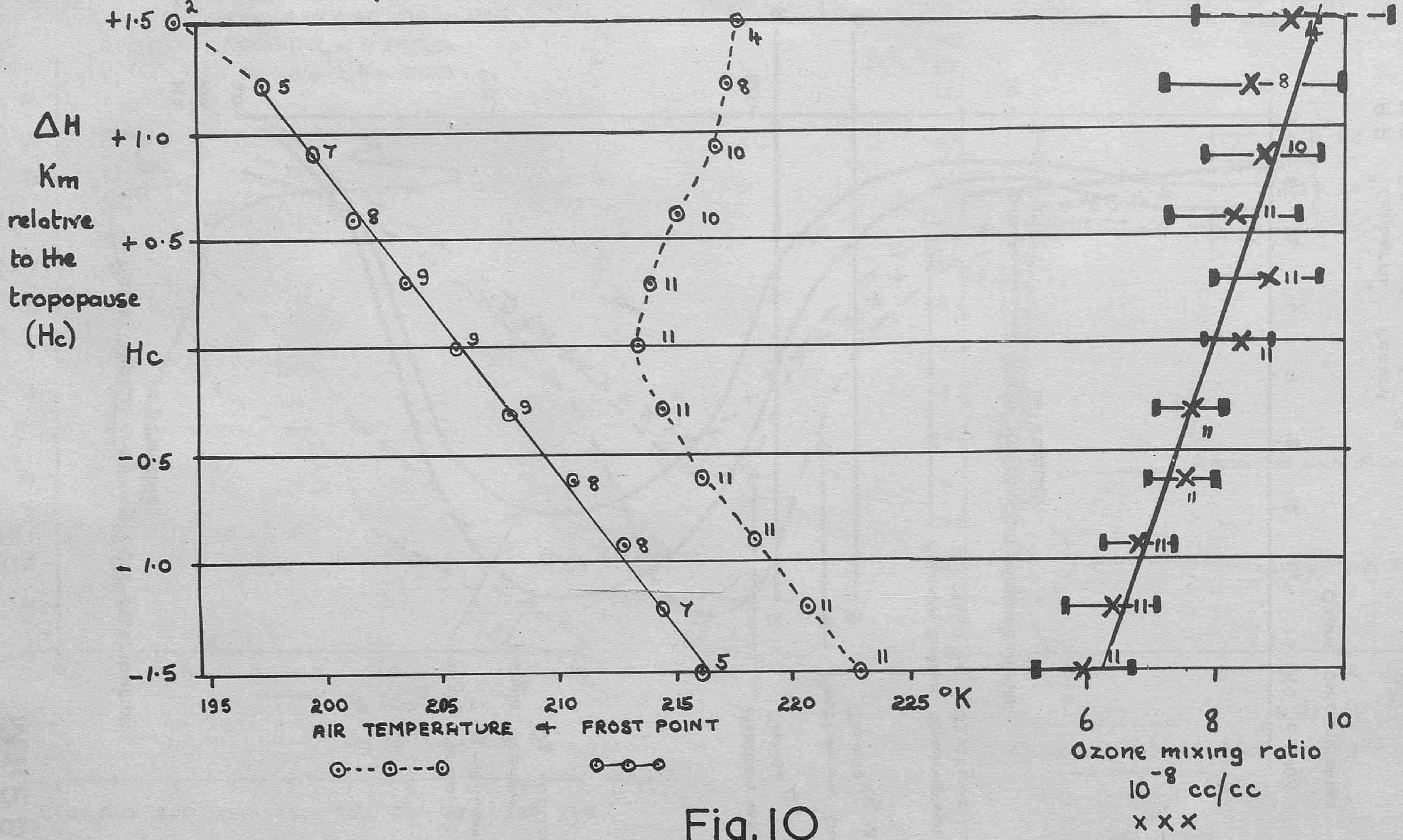


Fig. 10

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M.R.P. 818
S.C. III/155
30th May, 1953.

AD 223061

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

# The temperature decay law of a naturally convected air stream with distance from its source of heat

by W. Railston
Royal Military College of Science, Shrivenham.

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## Abstract

Temperature measurements have been made both vertically and transversely in the plume of air rising by natural convection from an electrically heated gauze, in a still atmosphere. Schlieren methods made the airstream visible, so that the type of air flow could be observed at different levels. The results are in good agreement with the dynamical theory of naturally convected gases developed by O.G. Sutton.

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## 1. Introduction

The mathematical analysis of convection [illegible] it was desired to make careful experimental measurements of the way in which the mean temperature of a hot air stream, rising under natural convection, [illegible] with increasing distance from the heat source [illegible]. The maximum measured rise of temperature [illegible] the hot air and air [illegible].

The plan of the research [illegible] methods of producing a uniform [illegible] at a number of points across the jet, at each of many different [illegible] heat source. [illegible] theoretical analysis of [illegible] source of heat, it was decided to use as small a source as [illegible] make a theoretical extension to a true point source. In this way [illegible] the mean temperature of the jet as a function of its height above [illegible] found.

Little prior experimental work had been done on this [illegible] (1941), who unfortunately gave few practical details of [illegible] source of heat, and omitted to state the rate of heat output [illegible] shown later that in making no attempt to render the air [illegible] a valuable means of observing the type of air flow and of [illegible] throughout the experiment.

It was further decided to use several different heat [illegible] so that a [illegible] could be made of the effect of the strength of the source on the temperature rise it caused at any given point.

## 2. Experimental procedure

MEAN VALUES of frost point, air temperature and ozone concentration relative to the tropopause

(No. of observations indicated beside each point)

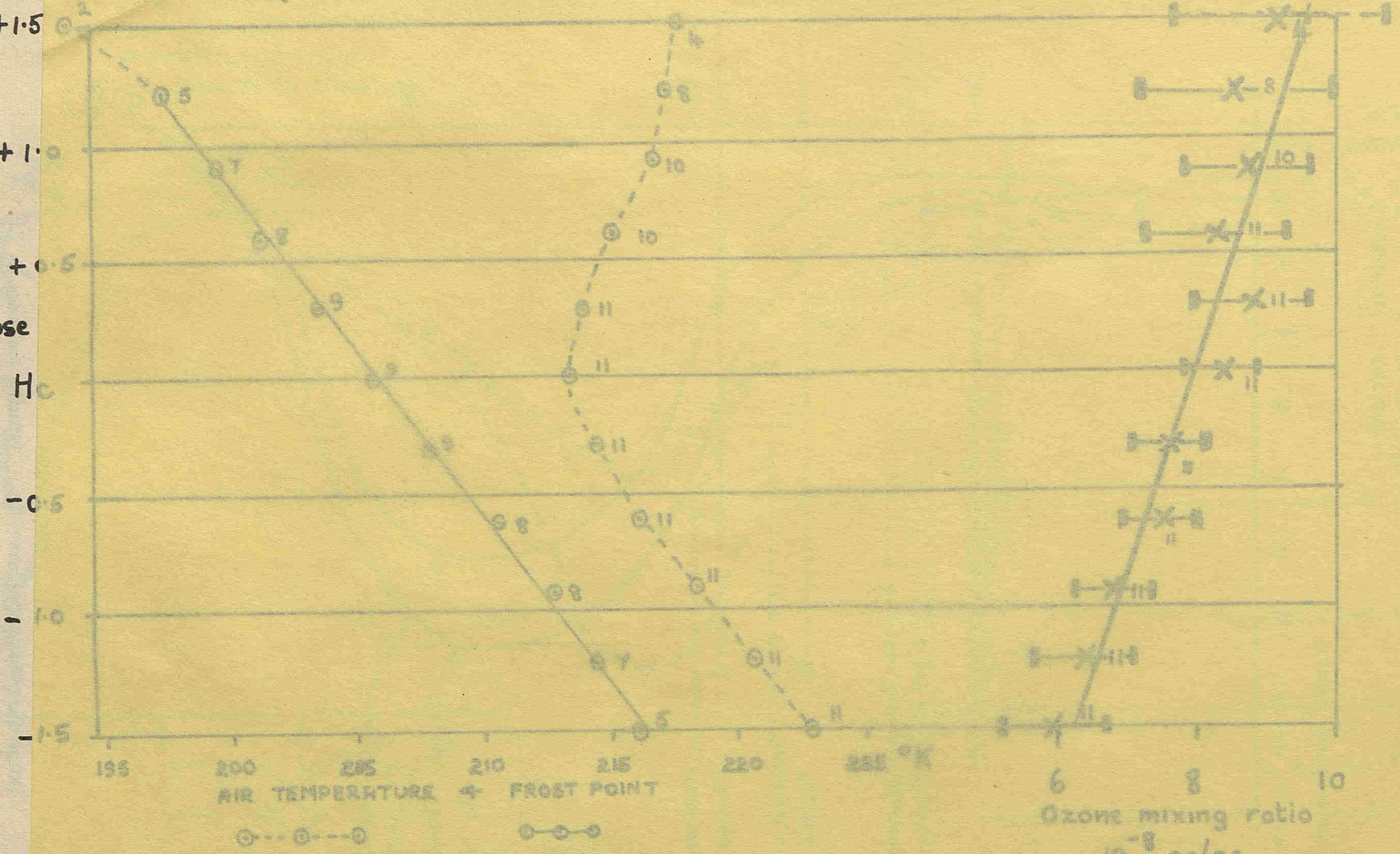

Fig.10

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AIR WEATHER SERVICE
TECHNICAL REFERENCE FILES

AD 223061

## AIR MINISTRY

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### Abstract

Temperature measurements have been made both vertically and transversely in the plume of air rising by natural convection from an electrically heated gauze, in a still atmosphere. Schlieren methods made the airstream visible, so that the type of air flow could be observed at different levels. The results are in good agreement with the dynamical theory of naturally convected gases developed by O.G. Sutton.

### 1. Introduction

The mathematical analysis of convection problems being notoriously difficult, it was desired to make careful experimental measurements of the way in which the mean temperature of a hot air stream, rising under natural convection, falls off with increasing distance from the heat source, when in a still atmosphere. The maximum measured rise of temperature was kept small ($< 21^{\circ}$C) so that the density of the hot air and its surroundings could be considered sensibly constant.

The plan of the research divided itself into the discovery and use of the best methods of producing a uniform, stable hot air jet, and of measuring the temperature at a number of points across the jet, at each of many different distances above the heat source. Since theoretical analysis of this problem usually refers to a point source of heat, it was decided to use as small a source as practicable and then to make a theoretical extension to a true point source. In this way, the variation of the mean temperature of the jet as a function of its height above a point source was found.

Little prior experimental work had been done on this problem, except by Schmidt (1941), who unfortunately gave few practical details of his work, especially of the source of heat, and omitted to state the rate of heat output from it. It will be shown later that in making no attempt to render the air stream visible, he overlooked a valuable means of observing the type of air flow and of ensuring similar conditions throughout the experiment.

It was further decided to use several different heat outputs, so that a test could be made of the effect of the strength of the source on the temperature rise it caused at any given point.

### 2. Experimental procedure

Trials were made of different types of electrically heated resistance, aiming to produce a uniform, stable air jet. Nichrome wire was at first wrapped into a "coiled coil" (as used in some electric lamps), of about 2 cm. major diameter and 2 mm. minor diameter. This clearly gave no heat directly to the central axis of the

/air

air jet, so it was replaced by another piece of nichrome wire wound closely in a zig-zag pattern between two opposite sides of a hollow square bakelite frame. The best source used was a piece of fine mesh wire gauze, originally produced for the electrical heating of flying suits. A piece of copper wire was soldered to each of two opposite sides of a square of the gauze of 3.5 cm. side. Alternating current flowed between these two sides, supplied from a variable auto-transformer receiving power from the mains through a constant voltage transformer, to minimize mains voltage fluctuations. The currents passed through the heater, together with the values of its resistance at the different temperatures they produced, were each measured on a precision potentiometer. Thus the consumption of the heater in watts was found for each current.

For temperature measurement at different points in the air jet, the platinum resistance thermometer, the thermocouple and the thermistor were each considered. The first was discarded, since although special methods allow the use of only a few millimeters of platinum wire, it is still not suitable for measuring temperature at a point.

Copper-constantan thermocouples were used in some of the initial measurements, but some changes of calibration were noted. As the hot junctions were being frequently moved in the air jet, these may have been due to mechanical changes in the leads caused by bending or kinking. Because of this and for greater convenience, thermistors were substituted for thermocouples. Thermistors themselves suffer from change of calibration, the resistance at a given temperature at first increasing slowly with time. This change was avoided by using thermistors which were at least a year old and which had reached a stable calibration by natural ageing. Artificial ageing, by heating to the maximum temperature to be measured for about 24 hours, was also successfully used.

The connection between the resistance R of a thermistor and its absolute temperature T was originally derived theoretically in the form $R = a \exp. (b/T)$ from considerations of the free energy of an electron gas. This was shown to differ considerably from experiment and was replaced by the equation $\log R = \log A - C \log T + B/2.303\ T$ by Becker, Green and Pearson (1947) who had noted that the graph of log R against $1/T$ increased linearly in slope with temperature. Bosson, Gutmann and Simmons (1950) have shown, however, that the resistance of a thermistor, at a given temperature, calculated from this formula, can differ by as much as 16.6 per cent from the experimental value, and have suggested the substitution of $\log R = A + B/(T+\alpha)$ which gives a maximum discrepancy of about 2.5 per cent.

In view of the uncertainties introduced by these formulae, the thermistors used were calibrated throughout the required temperature range against standard N.P.L. certificated thermometers, making use also of the ice point, the steam point and the transition point of sodium sulphate decahydrate, whose value (32.38°C) is accurately known. It being necessary to calibrate the thermistor in the same medium as it was to be used, it was placed together with a stirrer and the standard thermometer in a thin-walled glass tube, which was then partially immersed in an oil bath which was electrically heated and stirred. For measurements at the standard temperatures, the thin-walled tube was placed in contact with the standard substance in a thermos vessel, and left for a long period to attain temperature equilibrium.

Resistance measurements were made by having the thermistor as one arm of a precision Wheatstone bridge having, as nearly as possible, equal arms. The criteria of Callendar (1910) for maximum bridge sensitivity were observed. The thermistors used were of the Standard Telephones type F 2311/300 having a dissipation constant of 0.62 milliwatt/°C in air. The bridge was energised from an accumulator through a potential divider arranged so that the power consumed by the thermistor did not exceed 4 microwatts and did not raise its temperature by more than about .006°C. To secure good sensitivity while using this very small power input to the bridge, a mirror galvanometer was used as balance indicator.

/At

At a temperature of 35°C, which was commonly used, this type of thermistor had a temperature coefficient of resistance of .03 per °C, which is eight times that of platinum. The heat sensitive bead was about 1 mm. in diameter and together with its glass mounting weighed 1.8gm., having a thermal capacity of only 1.4 joules per °C. The thermistor and its leads were supported in a rigid holder so that the whole was free from stress. This is essential if results are to be reproducible. Two similar thermistors were used, one to measure the ambient temperature and the other the temperature at different points in the air jet. They were attached to the compound slide taken from a watchmaker's lathe, capable of horizontal motion in two directions at right angles. The slide could be raised or lowered on a stout steel rod fixed to a heavy base. The position of the thermistor in the air jet was determined by means of a cathetometer.

The experiment was first carried out in a small room having light-proof blinds over the windows. These blinds were also good draught excluders and it was because of this that they were drawn. Even on days of little wind, however, it was found that the temperature at a fixed point in the air jet fluctuated continually in a chance manner, the jet acting as a very sensitive draught detector. The source of heat and the thermistors were then placed in a cupboard, open on one side. A hole about 10 cm. in diameter was placed in the roof of the cupboard vertically above the heat source, so that the hot air could escape without setting up a convective circulation in the observation chamber. This system reduced draughts, but they were still troublesome. It became obvious that they would have to be further reduced and the flow of hot air in the jet made visible. The extent of the influence of draughts would then be observed clearly, together with the efficacy of attempts to remove them. Further, the type of flow, whether laminar or turbulent, could be decided at once. The change in refractive index caused by heating the air was extremely small since the maximum heat output was only 20 watts. For the first trial, a shadowgraph method was used, light from a motor-car headlamp bulb with a short filament, slightly over-run, passing through the air jet and striking a white screen. Such a system shows changes in illumination caused by the second derivative, $\partial^2\rho/\partial z^2$ , of the density $\rho$ with respect to distance $\underline{z}$ in the direction of flow. Hence sharp shadow images will only be produced where there are large non-linear changes in density as in the case of a shock wave. It was not surprising therefore that the method failed to delineate the path of this hot air stream at distances of more than about 20 cm. from the weak heat source. A method of greater sensitivity was required.

On the request of Dr. O.G. Sutton to the Meteorological Research Committee, a high quality concave spherical mirror of 16 in. aperture and 200 in. radius of curvature was made available, and used to construct an efficient "Schlieren" system. The single mirror arrangement was selected for reasons of economy, in view of the large field of view desired.

Light from a mercury, high pressure, compact source lamp, A, was passed (fig. 1) through a condensing system consisting of two good quality camera lenses B, C, of 9 cm. aperture, with an iris diaphragm D placed between them, to produce an intense source, about 1 mm. in diameter, at E, which was the centre of curvature of the high grade spherical mirror FG. The mirror was slightly turned about a vertical axis, so that the reflected light was brought to a focus at K, instead of returning to E. The knife edge, O, could be made to block the path of the light at K, if required. It was usually used vertically, but could also be made horizontal. It could be moved across the light rays by a micrometer screw slide. The usual separation of K from E was about 2 cm. and so since the distance from E to the mirror exceeded 500 cm. the incident and reflected rays were very nearly coincident, especially near the mirror. In fig. 1 the amount of the separation E K was magnified to allow the incident and reflected rays near the mirror to be shown clearly. Being a mirror system, it had the advantage over a lens of freedom from chromatic aberration and the effects of inhomogeneity in the thickness of the glass. Further since the incident and reflected rays were nearly coincident, astigmatism was negligible and the virtual image, behind the mirror, was exactly masked by the object, (the hot air stream), because object and image subtended the same angle at the centre of curvature. It will be readily seen that since the rays traversed the hot air twice, and were twice deflected in the same direction, a magnification of deflection resulted. This principle is very useful to increase the minute angular deflections produced by small amounts of heat, since these cause little departure from the coincidence of incident and reflected rays near the mirror. (With the steep density gradients encountered in the shock waves from projectiles, however, it is likely that the outward and returning rays would traverse markedly different paths and a double image would result).

Any light not stopped by the knife edge, passed on to the prism L where it was deviated so that it fell on the ground glass screen M N, which could be replaced by a camera when desired. As the knife edge was moved forward to obscure more of the image at K, a uniform darkening of the screen M N was caused, provided the knife edge was at exactly the same distance from the screen as K, since every point of K received light from every point of the mirror. If the knife edge was traversed across the reflected rays at a greater or less distance than K from the mirror, its shadow could be seen to travel across the screen, causing non-uniform darkening.

This uniform darkening produced with the knife edge at O occurred only if the medium in front of the mirror was of uniform density. Now suppose the space H J was of constant density except for a small region where there was a density gradient. This density gradient would cause a gradient of refractive index, making the small region act like a prism. The light passing through it would be deviated as compared with that through the surrounding space. If it was deviated towards or away from the knife edge, the latter would allow less or more light to pass on to the prism and so to the screen. Consequently the image of this small region would appear darker or brighter than the surrounding area where no density gradient existed. When the hot air was moving, density gradients $\partial\rho/\partial z$ and $\partial\rho/\partial y$ existed through the plume, having different values at different points. (y̲ was the direction at right angles to the direction of flow, z̲, and also to the axes of the optical system.) The resultants of these gradient components therefore produced brighter or darker regions in the image on the screen than the uniform light intensity which existed before the air was set in motion. If the small region had a constant density gradient in it, all rays passing through it would be equally deviated and would produce a corresponding area of uniform light intensity as its image on the screen. Since the knife edge acted as a stop due to rays being deflected at right angles to its edge, it was in this perpendicular direction that the system was most sensitive. Thus with the knife edge vertical, the horizontal components $\partial\rho/\partial y$ of density gradients could be detected most readily. When the knife edge was horizontal, the system was most sensitive to vertical density gradients $\partial\rho/\partial z$. In general, a density gradient which caused light to be deviated along the knife edge caused no change in the illumination of the screen and so could not be detected.

Somewhat greater sensitivity was obtained when a narrow illuminated slit, parallel to the knife edge, was used as the light source. This, however, caused inconvenience in having to be re-orientated with the knife edge when the latter was turned to detect horizontal instead of vertical density gradients. The need for this rotation of the source was avoided by the use of a very small circular hole (a nearly closed iris diaphragm) as source. The system was still very sensitive, for it would show warm air rising from the hand. Air currents from a slowly cooling thermometer could be detected when it showed a temperature only 1°C above the surroundings. The change in refractive index of air heated through 1°C is only 1 part in a million. The long radius of curvature of the mirror played a large part in this high sensitivity. By the use of white light, even greater sensitivity can be obtained as in addition to changes of light intensity there are gradations of colour, which are extremely sensitive to change in density gradient. The reason for the presence of colours is due to their differing deviations in the air jet causing different degrees of obscuration by the knife edge. The mercury, high pressure lamp, however, was preferred because its high intensity allowed exposure times to be reduced when air streams were photographed, giving greatly improved definition. Before the heated gauze, source of the air jet, was switched on, the knife edge was adjusted to give a faint uniform illumination over the image of the mirror in the screen. In this faint light, the thermistors could be seen outlined by diffraction bands. When the air jet was started, the thermistor could be seen in it and the type of air flow observed. The thermistor had initially been placed vertically above the centre of the heating gauze, by using a plumb line. The Schlieren method showed whether the sensitive bead was truly central in the air stream and when the latter was disturbed by the slightest draught. Gawthrop (1931) has given some details of Schlieren systems and applications.

The Schlieren mirror and the heating gauze were now placed in a large wood box, approximately 1 metre by 1 metre by 2 metres tall. The heated air passed out of a hole in the top, whilst cool air entered through a large opening in the bottom, passing through two screens of perforated zinc to remove eddies before arriving at the gauze. The front of the box had two half doors, one above the other, to give access. The upper door had a large aperture covered by a sheet of selected plate glass. With this door open, light could fall directly on the mirror and adjustments could be made inside the box. With the door closed to exclude draughts, the image on the screen showed dark horizontal bars about 2 cm. wide, with clear spaces about 3 cm. wide between them. These were due to regular changes in refractive index caused by the rollers used in the manufacture of the plate glass. It was easy to arrange that the important region round the thermistor appeared in a clear space. Further, the dark bars could be lightened by reducing the Schlieren sensitivity. On windless nights, it was often possible to work with this door open. By contrast, at times of strong winds, with the door closed, and with closely fitting blinds over the windows the air jet would be too disturbed by draughts and mechanical vibration to make dependable measurements. On hot, sunny days, even though windless, one wall of the laboratory would become warm, causing a disturbing circulation of the air in the room. All these disturbances were made visible by the Schlieren method.

The measurement procedure was to place the jet thermistor at the required height above the gauze, as measured by the cathetometer, and to check its position vertically above the centre of the gauze by means of the plumb line. The air stream was then observed visually by the Schlieren method, and conditions of steady flow awaited. When these were attained, the jet and ambient temperatures were measured, using their respective thermistors. The temperature excess of the jet over its surroundings was thus determined. The temperature measurements were repeated several times and the spread of the results for the temperature excess observed. If this was small, the average temperature excess was accepted, but if not a greater number of readings was taken or the experiment stopped until better conditions prevailed. The thermistors were then raised to a new position and the procedure repeated. The temperature excess on the central axis of the jet was measured from 5 to 70 cm. above the heating gauze, at intervals of 5 cm. In addition, at 20, 40 and 60 cm. above the gauze the thermistor was traversed across the jet from the central axis to beyond a point where the temperature excess fell to one-tenth of that of the axis. This was to test the law of lateral temperature decay in the jet. It was done for one heat output only.

In all these cases, heat was conveyed from the gauze to the thermistor by convection, radiation and conduction. It was considered that the part played by the last was negligible, as all air temperatures measured were entirely outside the boundary layer round the gauze. The effect of radiation, however, had to be determined. This was done at the different distances from gauze to thermistor where the temperatures due to the combined effects of convection and radiation had already been measured. The method was to use exactly the same conditions, except that the thermistor was placed at the same distance below the gauze as it had previously been above. In this way the temperature excess due to radiation alone was determined. The measurements were discontinued when distance from the gauze caused the temperature excess to be less than 0.1°C.

It may be asked why the horizontal heating gauze used was not made smaller than 3.5 cm. square so as to make the theoretical extension to the case of a point source easier. This was because, with a smaller heater, the region of maximum temperature on the axis of the jet was very small, so that a transverse movement of the thermistor of only 1 mm. might cause its reading to change by several degrees when near the gauze. In short, the location and measurement of the point of maximum temperature became more difficult, and increasingly liable to inaccuracy as the area of the gauze was decreased.

/3.

## 3. Results

The results for the excess of temperature of the central axis of the jet above its surroundings at various temperatures from the heating gauze are given in table 1, for three different heat outputs. They have been corrected for the effect of radiation, and represent the contribution of convection alone. The heat outputs used were 4.87, 9.92 and 20.37 watts. The results are graphed in fig. 2 and the effects of radiation alone can be seen in fig. 3. The ambient temperatures at which these results were taken were all between 10.4 and 16.0°C.

Table 1. Points on central axis of jet

| | Heater output (watts) | | |
|---|---|---|---|
| | 4.87 | 9.92 | 20.37 |
| Height (z) of thermistor above heater (cm.) | Excess temperature $\Delta\theta$ of jet axis above surroundings (°C) | | |
| 5.0 | 7.8 | 12.9 | 20.5 |
| 10.0 | 5.2 | 7.7 | 12.2 |
| 15.0 | 3.4 | 5.8 | 9.2 |
| 20.0 | 2.7 | 4.7 | 7.2 |
| 25.0 | 2.1 | 3.4 | 5.4 |
| 30.0 | 1.6 | 2.8 | 4.3 |
| 35.0 | 1.4 | 2.3 | 3.8 |
| 40.0 | 1.3 | 2.1 | 3.1 |
| 45.0 | 1.1 | 1.7 | 2.6 |
| 50.0 | 0.9 | 1.4 | 2.1 |
| 55.0 | 0.8 | 1.3 | 2.0 |
| 60.0 | 0.8 | 1.1 | 1.7 |
| 65.0 | 0.7 | 1.0 | 1.4 |
| 70.0 | 0.6 | 0.9 | 1.2 |

In fig. 2 both $\Delta\theta$, the excess temperature on the axis of the jet, and $\underline{z}$, the vertical distance of the point from the gauze, are plotted on logarithmic scales. Since the resultant graph is curved, its equation is clearly not given by $\Delta\theta = c/z^n$, $\underline{c}$ being a constant. An attempt was therefore made to find a theoretical point source, at a distance $\underline{a}$ below the heating gauze, from which the airstream could be assumed to rise conically. If this was correct the $\Delta\theta - z$ relation would be $\Delta\theta = c/(z+a)^n$. An approximate value of $\underline{a}$ was found by selecting three typical points on each of the curves of fig. 2 in turn. For one curve let the co-ordinates of the three points be $(\Delta\theta_1, z_1)$, $(\Delta\theta_2, z_2)$ and $(\Delta\theta_3, z_3)$.

Then

$$\frac{\log \Delta\theta_1 - \log \Delta\theta_2}{\log \Delta\theta_1 - \log \Delta\theta_3} = \frac{\log (z_2 + a) - \log (z_1 + a)}{\log (z_3 + a) - \log (z_1 + a)} .$$

Let the first fraction be called P and the second S. Inserting the numerical results for $\Delta\theta$ , the value of P could be found. A number of trial values of $a$ were inserted in S and so numerical values of S were calculated, care being taken to use only those values of $a$ which gave small values (positive or negative) of S - P. The graph of S-P against $a$ was found to be linear (fig. 4) and the required value of $a$ which made S-P = 0 was read from it. This was done for each of the curves of fig. 2, and the average of the three values of $a$ found to be 10.2 cm.

Still retaining the same meaning for $z$, the distance of any point on the axis of the jet from the theoretical point source was then $(z + 10.2)$ cm. A new graph of $\Delta\theta$ against $z + 10.2$ was then drawn (fig. 5). The three resultant curves are seen to be linear, except for the last five points at 20.37 watts, and the last three at 9.92 watts. The Schlieren system showed the air flow at these points, distant from the source, to be completely turbulent. The flow for all points falling on the linear curves was shown to be laminar or only slightly turbulent. From the slope of these curves the value of $n$ was found to be 1.52 (to two decimal places). Writing the heat output as Q watts, the above equation was then rewritten as $\Delta\theta = k\, Q^x/(z + 10.2)^{1.52}$ , $k$ and $x$ being constants.

It was desired to find the relation between the temperature on the central axis and the mean temperature across the jet at the same height. Temperatures across the jet were measured by traversing the thermistor by means of the compound slide to a point beyond where the temperature fell to one-tenth of the axis value. Results, taken at intervals of 1 cm. for a heat output of 20.37 watts, are given in table 2, and graphed in fig. 6.

Table 2. Transverse excess temperatures across the jet due to convection only

| | Distance from jet axis (cm.) | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 | 15 | 16 |
| VERT. DIST. ABOVE HEATER (CM.) | EXCESS TEMPERATURES ABOVE SURROUNDINGS (°C) | | | | | | | | | | | | | | | |
| 20 | 7.0 | 6.2 | 4.7 | 3.3 | 2.3 | 1.5 | 0.9 | 0.5 | | | | | | | | |
| 40 | 3.1 | 2.9 | 2.7 | 2.4 | 2.0 | 1.6 | 1.2 | 1.0 | 0.7 | 0.5 | 0.4 | 0.2 | | | | |
| 60 | 1.7 | 1.6 | 1.6 | 1.4 | 1.3 | 1.1 | 1.0 | 0.8 | 0.7 | 0.6 | 0.5 | 0.4 | 0.3 | 0.2 | 0.1 | 0.1 |

The central axis excess temperatures, $\Delta\theta$ , for these heights above the heater are given in table 1. From fig. 6 the area under each curve, up to the point whose excess temperature was $\Delta\theta/10$ , was calculated by Simpson's rule. The radius $r$ of the jet, at each height explored, taken as the horizontal distance of this point from the axis, was also measured from the curve. Dividing the area by $r$, the mean excess temperature, $\overline{\Delta\theta}$ , across the jet was determined for each height. The mean of the results gave $\Delta\theta = 1.80\ \overline{\Delta\theta}$ , suggesting a transverse temperature distribution in the jet similar to the normal law of error.

It was now possible to convert the large number of axial excess temperatures $\Delta\theta$, already found experimentally, into mean excess temperatures, $\overline{\Delta\theta}$, across the jet. The equation giving the variation of $\Delta\theta$ with height above the theoretical point source also held for $\overline{\Delta\theta}$ if the constant k was changed to p, where $k = 1.80\ p$.

In fig. 7 are drawn a series of curves connecting $\Delta\theta$ and Q, the heat output in watts on logarithmic scales. From these curves, up to a height of 45.2 cm. above the theoretical source, the average value of x in the equation $\Delta\theta = kQ^x/(z+10.2)^{1.52}$ was found to be 0.675, from the slopes of the curves. The mean value of k, up to the same height, was 164.3, giving a corresponding value of p of 91.3.

The values of r obtained from fig. 6 are given in table 3 and graphed against distance from the theoretical source in fig. 8, using logarithmic scales for each variable. The curve is a straight line representing an equation of the type $r = A(z + 10.2)^B$. From the slope of the curve, $B = 0.85$ and from the intercept on the horizontal axis, A was calculated to be 0.40.

Table 3. Values of r extracted from fig. 6

| | | | |
|---|---|---|---|
| Radius of jet, r cm. | 7.3 | 11.3 | 15.0 |
| Height above theoretical source, $z_o$ cm. | 30.2 | 50.2 | 70.2 |

Hence if we call the height of any point in the jet above the theoretical point source, $z_o$, where $z_o = z + 10.2$, the final results of this series of experiments are contained in the equations $\overline{\Delta\theta} = 91.3\ Q^{0.675}\ z_o^{-1.52}$ and $r = 0.40\ z_o^{0.85}$. The value of $z_o$ in terms of z clearly applies only to the particular size and shape of heater used in these experiments, but it is believed that the other results are general.

## 4. Comparison of these results with theory

This problem has been treated theoretically by Schmidt (1941) and Sutton (1948 and 1950). Both assumed the mixing length concept in dealing with the formation of a vertical airstream, but whereas the first used rather laborious infinite series in his solutions, the second got more direct results by assuming a close similarity between the spread of a hot gas stream and the diffusion of a cold gas in the atmosphere, provided the temperature of the hot gas was not much greater than its surroundings. He pointed out that in adapting the diffusion analogy to the case of the dispersion of a hot gas stream, the velocity of the natural convection, which decayed to zero with the excess temperature of the jet, could be considered as the wind involved in the diffusion. The reason for using only a small rise of temperature was so that the density of the gas could be considered constant. In the experiments just described, the maximum temperature rise was 20.5°C, and a typical ambient temperature was 13°C, so the maximum density change was only 7 per cent.

For the variation of the radius r of the jet with $z_o$, Schmidt assumed a relation $r = \text{const.} \times z_o$, while Sutton calculated that $r = 1.5\ C\ z_o^{m/2}$ and, from the results of observations on smoke clouds in the atmosphere, he took $m = 1.75$, so obtaining the result $r = 1.5\ C\ z_o^{0.875}$, where C is a constant. If the present writer's experimental results for r and $z_o$ are inserted in the latter equation, the mean value obtained for C is 0.244. The writer found the power of $z_o$ to be 0.85. Using this result, the mean value for C is 0.269.

Sutton (1950) calculated a value for $\overline{\Delta\theta}$ which reduces to

$$\overline{\Delta\theta} = 5.94\ 10^{-3}\ T_a^{1/3}\ Q^{2/3}\ C^{-5/3}\ (c_p\rho)^{-2/3}\ z_o^{-1.46}$$

where $T_a$ is the absolute ambient temperature, $c_p$ the specific heat of the air at constant pressure and $\rho$ is its density. It will be noted that $\overline{\Delta\theta}$ is not sensitive to the value of $T_a$, $c_p$ or $\rho$. If we take $T_a = 290°K$, $c_p = 0.24$ and $\rho = 1.205\ 10^{-3}$ gm/cc at 20°C and 760 mm., then this equation can be simplified to

$$\overline{\Delta\theta} = 8.99\ C^{-5/3}\ Q^{2/3}\ z_o^{-1.46} \quad .$$

If we take C to be 0.244, we get $\overline{\Delta\theta} = 94.2\ Q^{2/3}\ z_o^{-1.46}$. Alternatively, if C is taken as 0.269, the result is $\overline{\Delta\theta} = 79.5\ Q^{2/3}\ z_o^{-1.46}$. In all these equations, Q is taken to be in watts, or joules per sec.

To test these formulae, table 4 was produced. In this the experimental results taken from table 1 are given (column D) alongside those calculated from $\overline{\Delta\theta} = 94.2\ Q^{2/3}\ z_o^{-1.46}$ (column E) and from $\overline{\Delta\theta} = 79.5\ Q^{2/3}\ z_o^{-1.46}$ (column F). To make comparison easy, the experimental values of $\Delta\theta$ in table 1 have been divided by 1.8 to give $\overline{\Delta\theta}$ in table 4, and 10.2 cm. have been added to $\underline{z}$ to give $z_o$.

The values for $\overline{\Delta\theta}$ are accurate to the first decimal place in each column. The experimental results are closely fitted by the formula $\overline{\Delta\theta} = 91.3\ Q^{0.675}\ z_o^{-1.52}$ taken from them and described towards the end of section 3.

Table 4. Comparison of experimental and calculated results

| Heater outputs (watts) | 4.87 | | | 9.92 | | | 20.37 | | |
|---|---|---|---|---|---|---|---|---|---|
| $z_o$ (cm.) | Mean excess temperature $\overline{\Delta\theta}$ across jet | | | | | | | | |
| | D | E | F | D | E | F | D | E | F |
| 15.2 | 4.3 | 5.1 | 4.3 | 7.2 | 8.2 | 6.9 | 11.4 | 13.2 | 11.2 |
| 20.2 | 2.9 | 3.4 | 2.8 | 4.3 | 5.4 | 4.6 | 6.8 | 8.7 | 7.4 |
| 25.2 | 1.9 | 2.4 | 2.0 | 3.2 | 3.9 | 3.3 | 5.1 | 6.3 | 5.3 |
| 30.2 | 1.5 | 1.9 | 1.6 | 2.7 | 3.0 | 2.5 | 4.0 | 4.9 | 4.2 |
| 35.2 | 1.2 | 1.5 | 1.3 | 1.9 | 2.4 | 2.0 | 3.0 | 3.9 | 3.3 |
| 40.2 | 0.9 | 1.2 | 1.0 | 1.6 | 2.0 | 1.7 | 2.4 | 3.2 | 2.7 |
| 45.2 | 0.8 | 1.0 | 0.9 | 1.3 | 1.7 | 1.4 | 2.1 | 2.7 | 2.3 |
| 50.2 | 0.7 | 0.9 | 0.8 | 1.2 | 1.4 | 1.2 | 1.7 | 2.3 | 1.9 |

Clearly, the theoretical formula of Sutton used in column F, with C as 0.269, gives the closest approach to the experimental results. Comparison with Schmidt's formula, $\Delta\theta = \text{const.} \times z_o^{-5/3}$, can only be made in terms of $z_o$, since he used only one, unstated value of Q, and his figures for $\overline{\Delta\theta}$ were only proportional to the actual values, the constant of proportionality not being given. It is clear that on these grounds Sutton's formula agrees better with the experimental results than Schmidt's. The experimental results give -1.52 as the power of $z_o$, while Sutton and Schmidt calculate -1.46 and -1.67 respectively.

/5

## 5. Conclusions

1. Sutton's formulae

$$\overline{\Delta\theta} = 5.94\ 10^{-3}\ T_a^{1/3}\ Q^{2/3}\ C^{-5/3}\ (c_p\rho)^{-2/3}\ z_\sigma^{-1.46}$$

and

$$r = 1.5\ C\ z_o^{0.875}$$

give very good agreement with experimental results when C = 0.269. With $T_a = 290^\circ K$, $c_p = 0.24$ and $\rho = 1.205\ 10^{-3}$ gm/cc these become

$$\overline{\Delta\theta} = 79.5\ Q^{2/3}\ z_o^{-1.46}$$

and

$$r = 0.40\ z_o^{0.875}$$

respectively.

2. The corresponding formulae derived directly from the experimental results are

$$\overline{\Delta\theta} = 91.3\ Q^{0.675}\ z_o^{-1.52}$$

and

$$r = 0.40\ z_o^{0.85}\ .$$

## Acknowledgements

Grateful acknowledgement is made to Professor O.G. Sutton, F.R.S., Dean of the Royal Military College of Science, and to A/Professor G.D. West for helpful discussions and for facilities provided. The Meteorological Research Committee is thanked for the loan of the Schlieren mirror. This paper is published by permission of the Commandant, Royal Military College of Science, Shrivenham.

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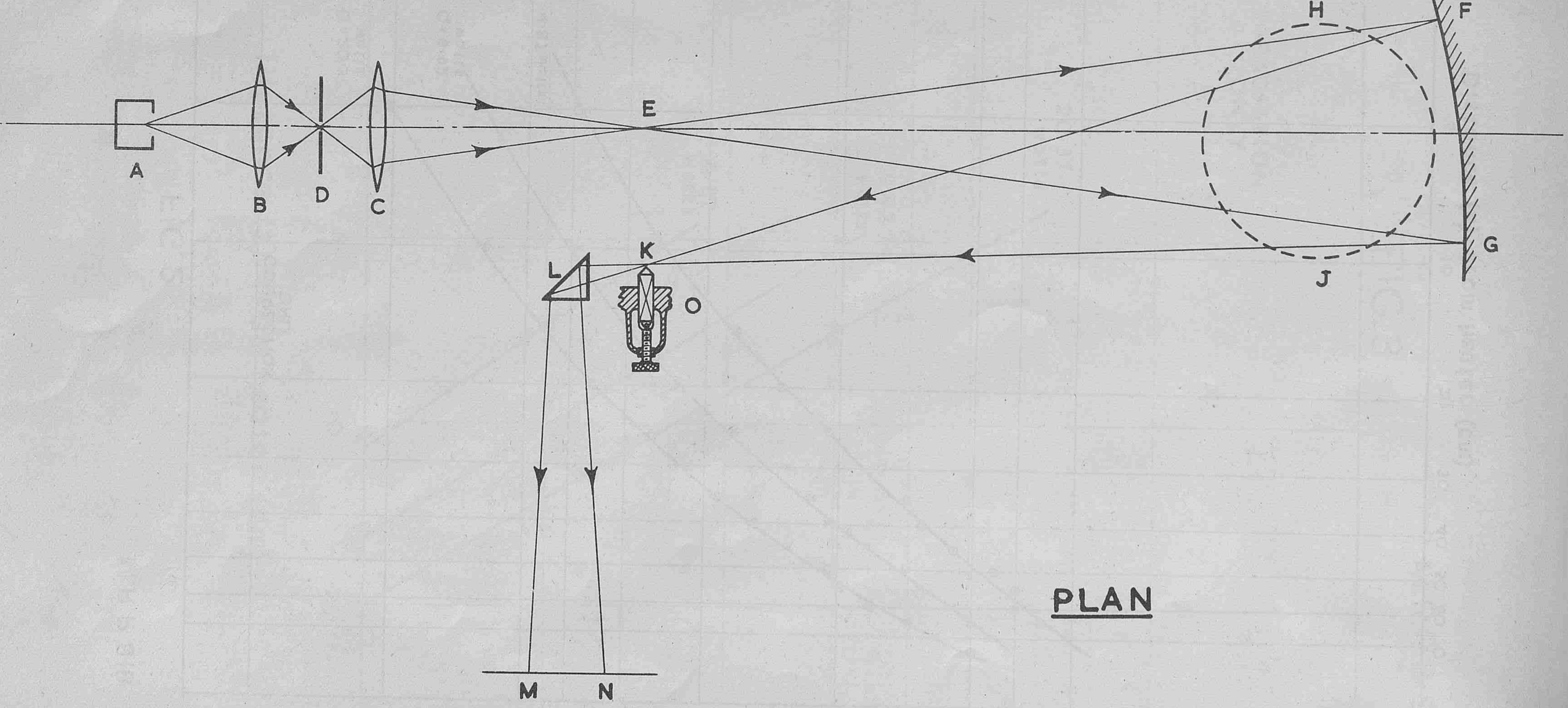
FIG. I


A
B
D
C
E
H
F
G
J
K
L
O
M
N

PLAN

FIG. 2

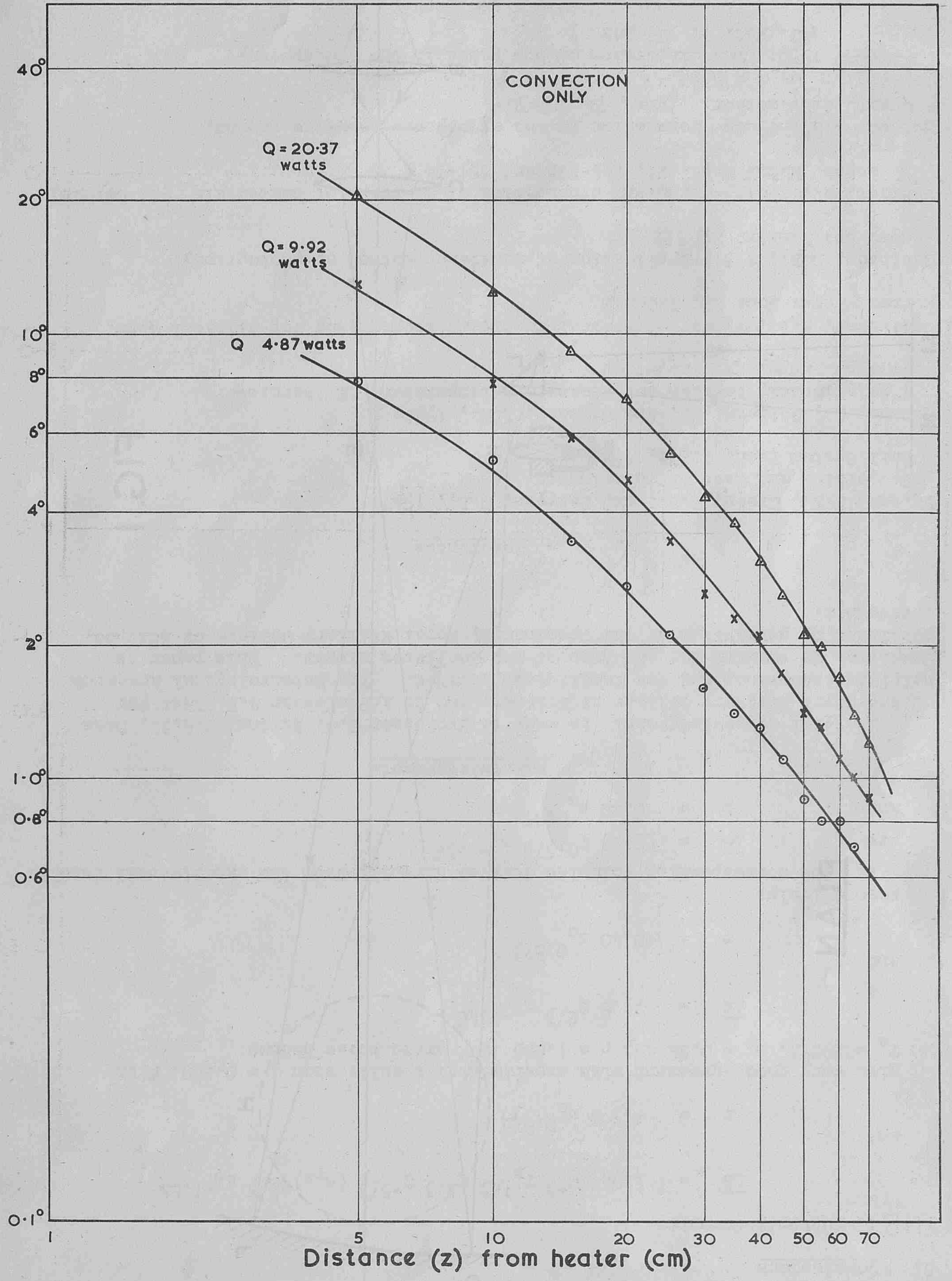
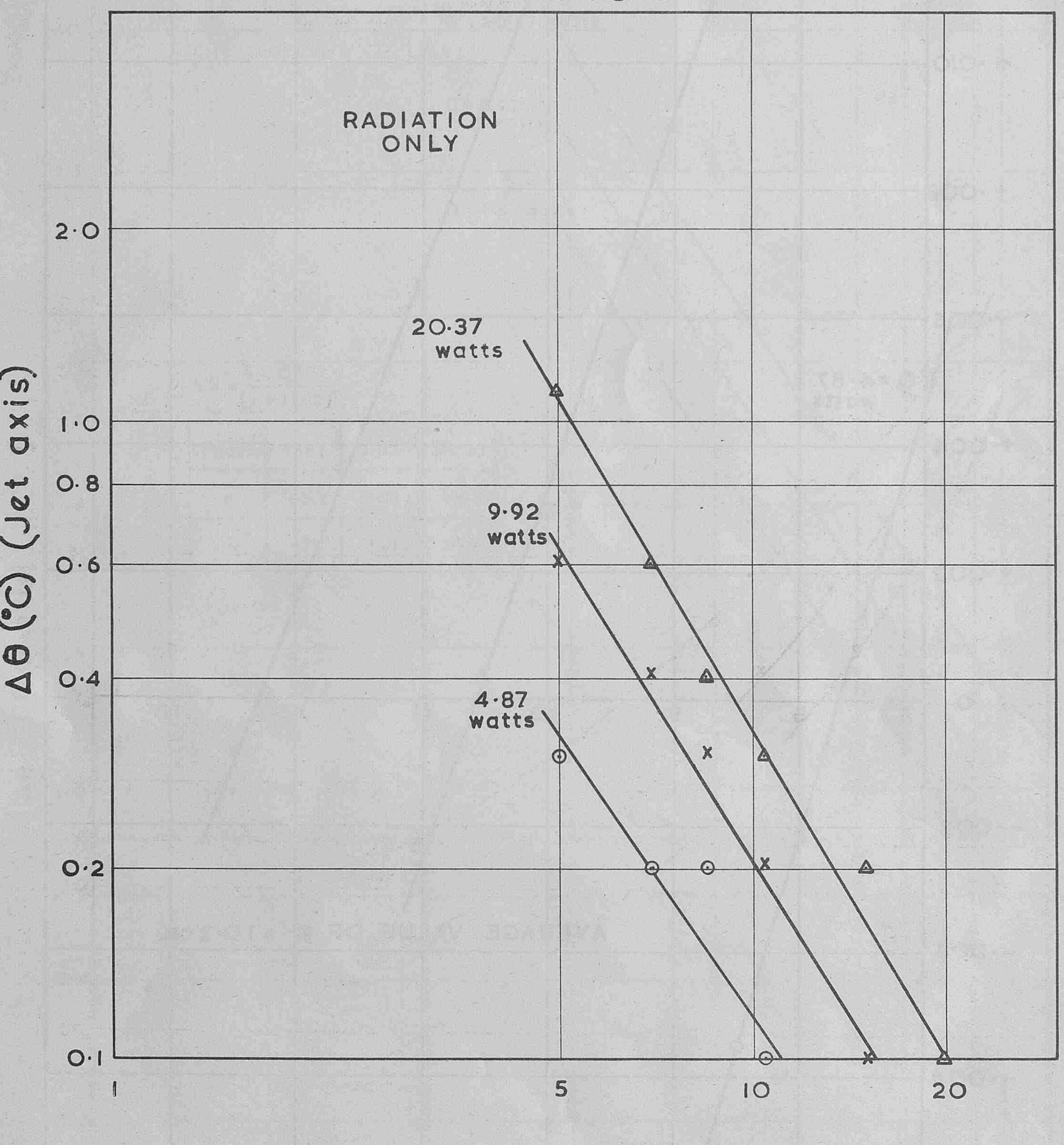
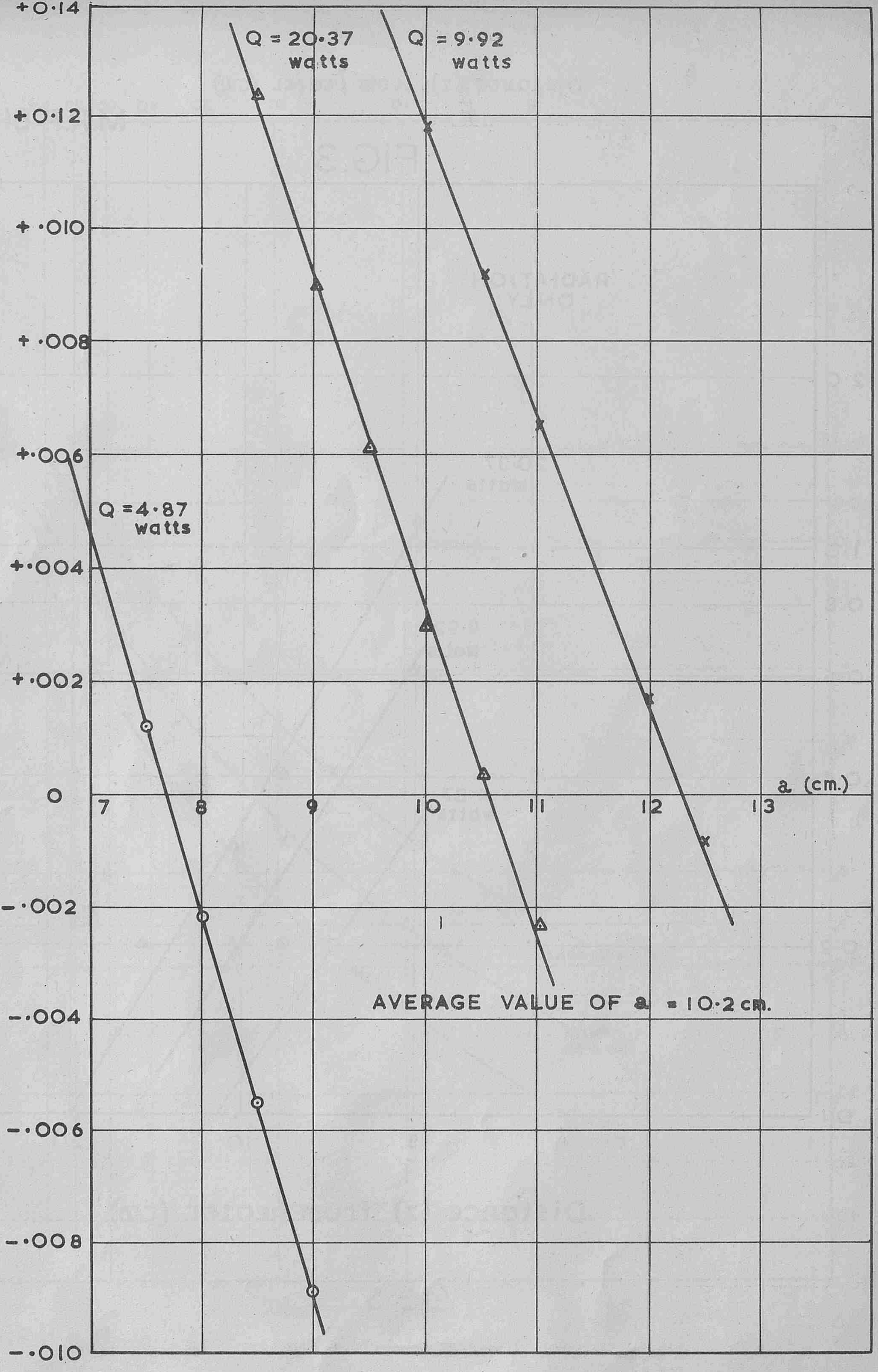

CONVECTION ONLY
Q = 20·37 watts
Q = 9·92 watts
Q 4·87 watts
40°
20°
10°
8°
6°
4°
2°
1·0°
0·8°
0·6°
0·1°
Δ θ (°C) (Jet axis)
1
5
10
20
30
40
50
60
70
Distance (z) from heater (cm)

FIG. 3



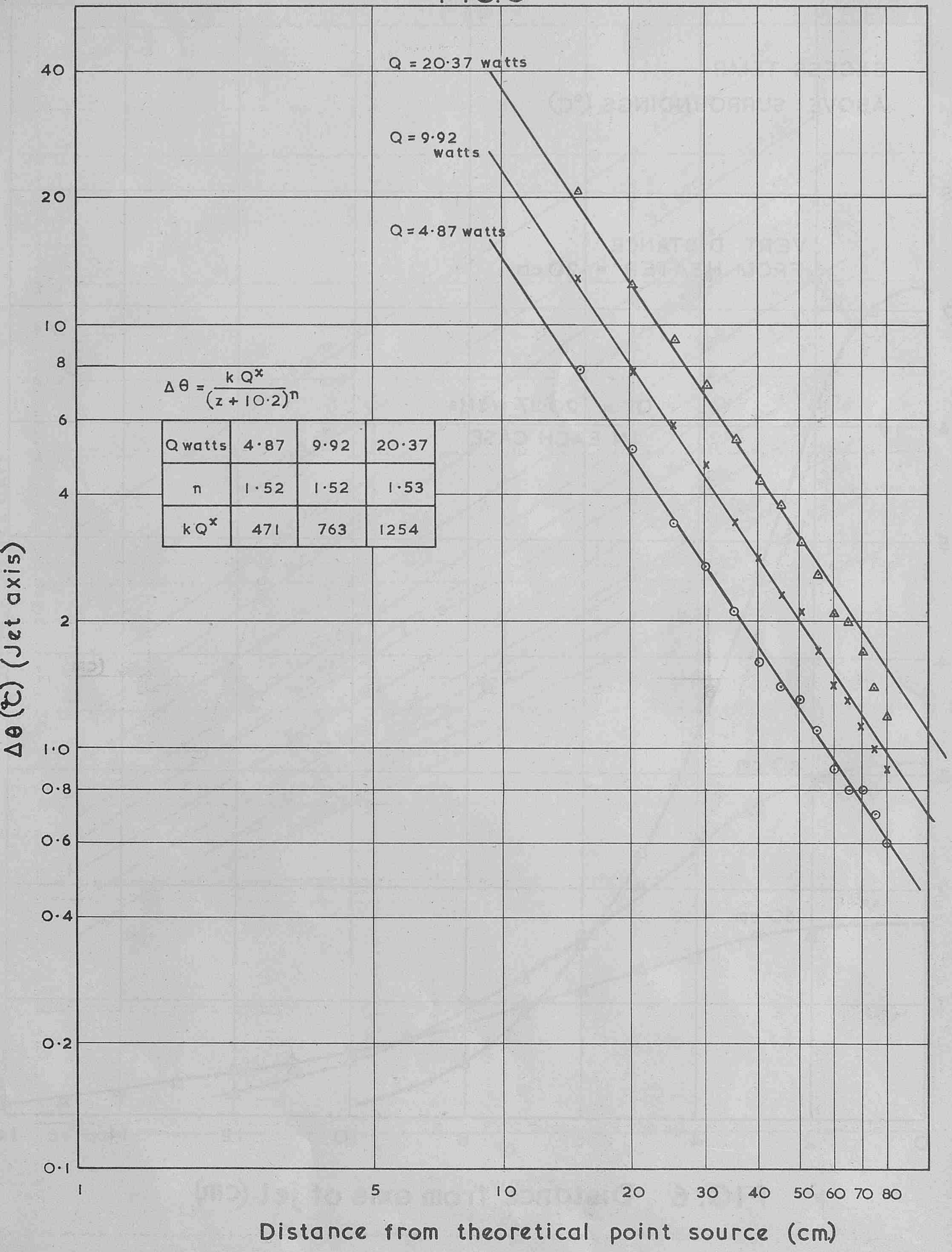
RADIATION ONLY
20·37 watts
9·92 watts
4·87 watts
2·0
1·0
0·8
0·6
0·4
0·2
0·1
1
5
10
20
Δθ (°C) (Jet axis)
Distance (z) from heater (cm)

FIG. 4



Determination of average value of a in $\Delta\theta = \frac{kQ^{x}}{(z+a)^{n}}$

FIG. 5



$$\Delta\theta = \frac{k\,Q^{x}}{(z + 10{\cdot}2)^{n}}$$

| Q watts | 4·87 | 9·92 | 20·37 |
|---|---|---|---|
| n | 1·52 | 1·52 | 1·53 |
| $kQ^{x}$ | 471 | 763 | 1254 |

FIG. 6

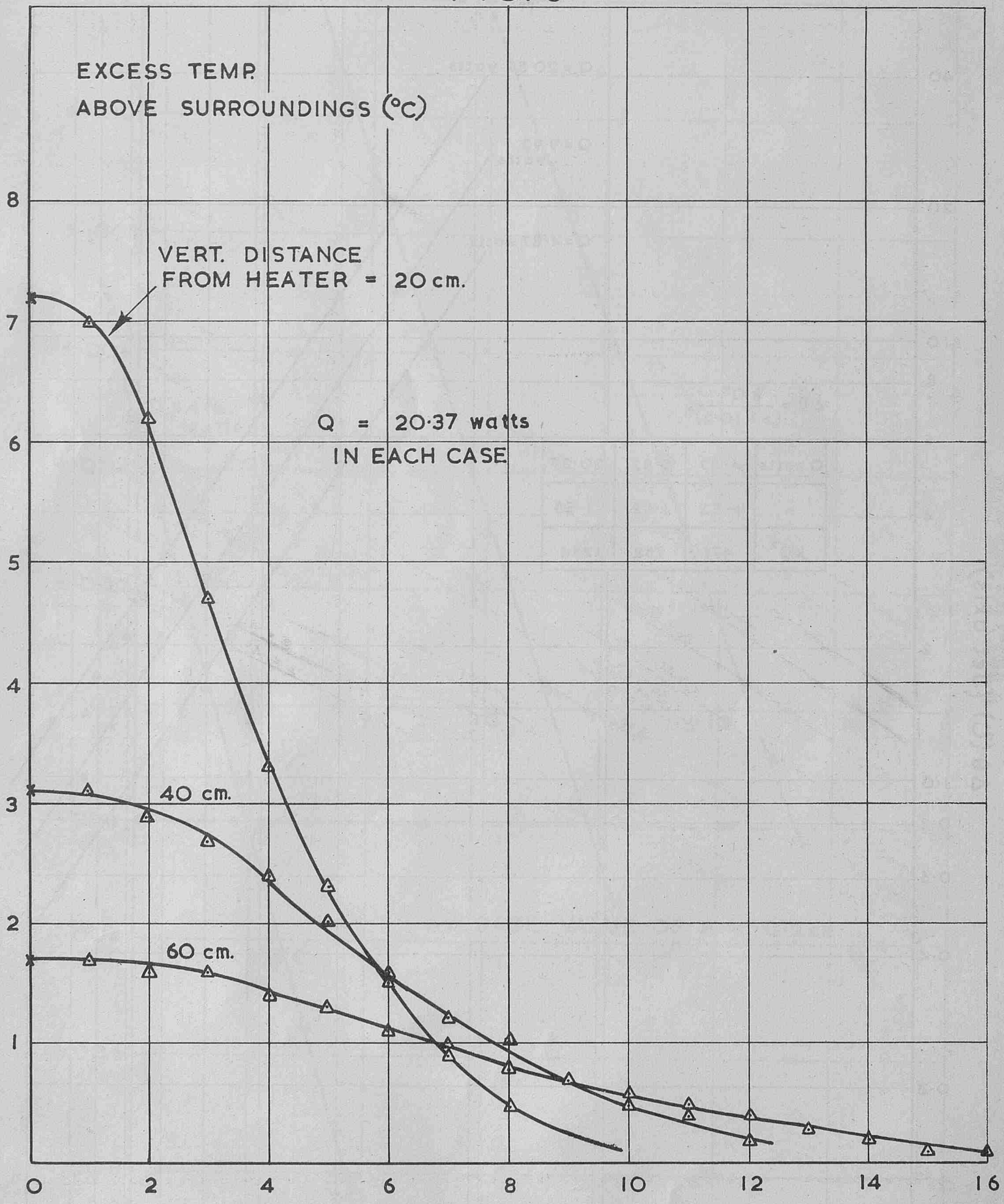


FIG. 6 Distance from axis of jet (cm)

# FIG. 7

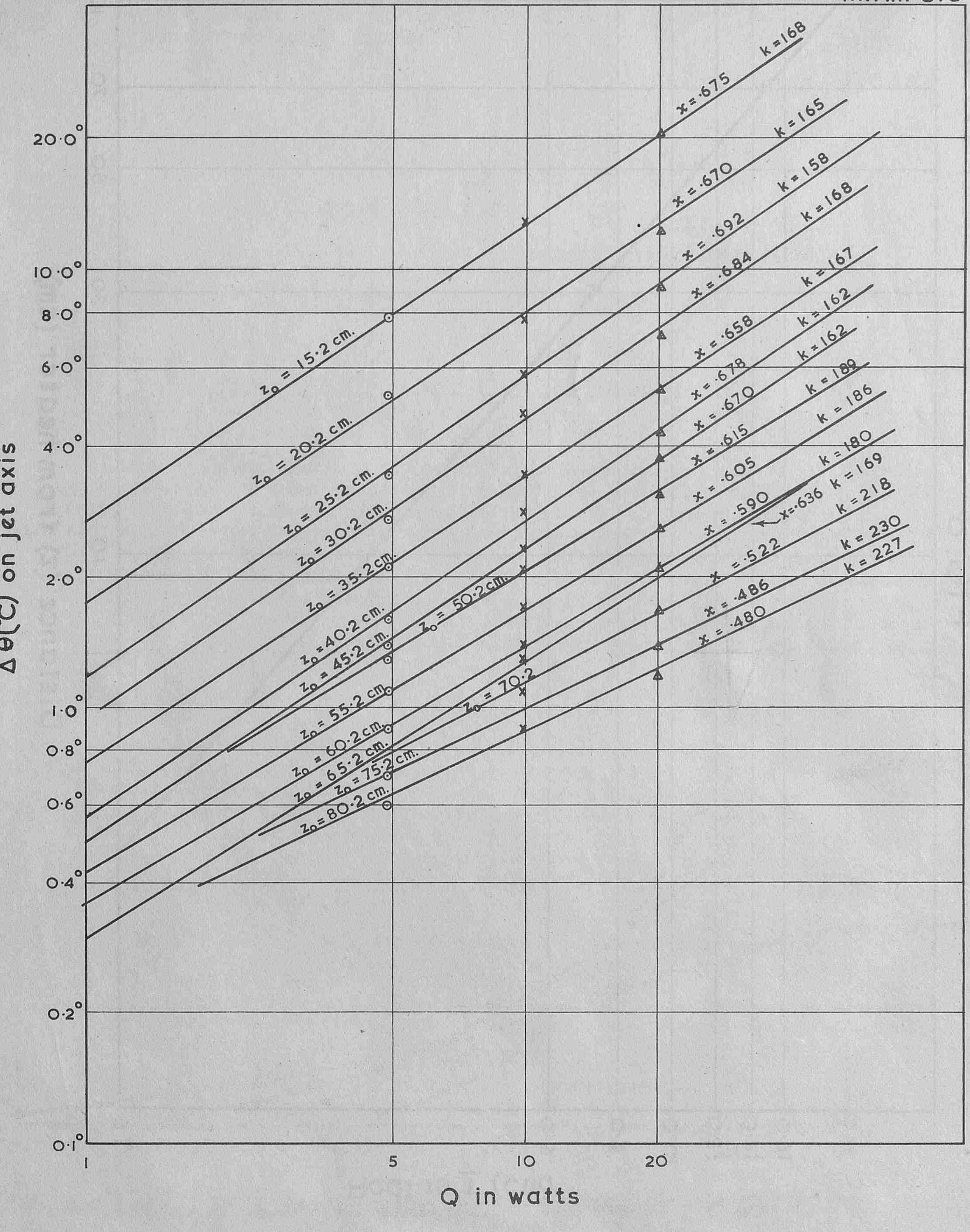
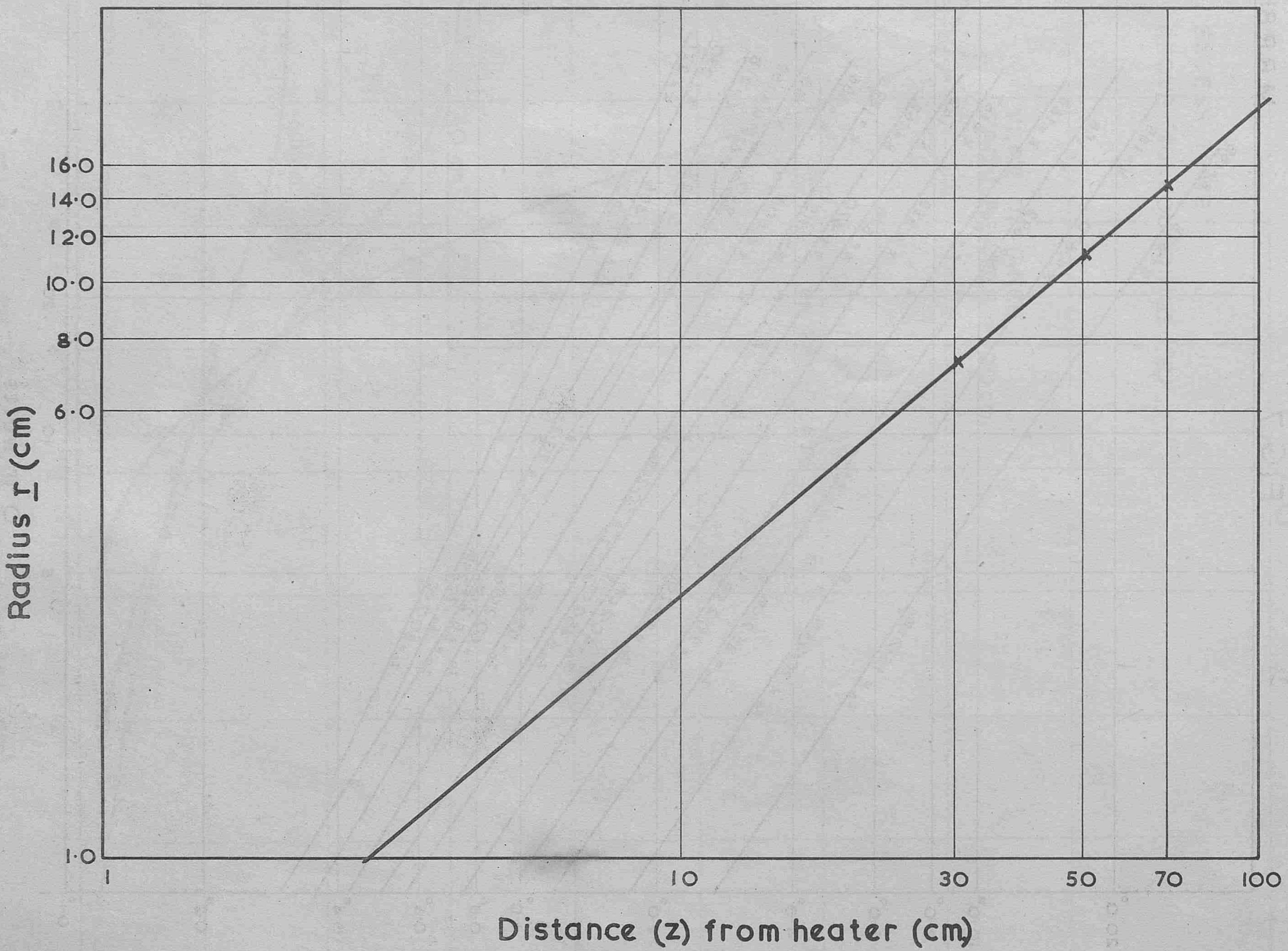

Δθ(°C) on jet axis
20·0°
10·0°
8·0°
6·0°
4·0°
2·0°
1·0°
0·8°
0·6°
0·4°
0·2°
0·1°
1
5
10
20
Q in watts
z₀ = 15·2 cm.
z₀ = 20·2 cm.
z₀ = 25·2 cm.
z₀ = 30·2 cm.
z₀ = 35·2 cm.
z₀ = 40·2 cm.
z₀ = 45·2 cm.
z₀ = 50·2 cm.
z₀ = 55·2 cm.
z₀ = 60·2 cm.
z₀ = 65·2 cm.
z₀ = 70·2
z₀ = 75·2 cm.
z₀ = 80·2 cm.
x = ·675 k = 168
x = ·670 k = 165
x = ·692 k = 158
x = ·684 k = 168
x = ·658 k = 167
x = ·678 k = 162
x = ·670 k = 162
x = ·615 k = 189
x = ·605 k = 186
x = ·590 k = 180
x = ·636 k = 169
x = ·522 k = 218
x = ·486 k = 230
x = ·480 k = 227

FIG. 8


Radius r (cm)
16·0
14·0
12·0
10·0
8·0
6·0
1·0
1
10
30
50
70
100
Distance (z) from heater (cm)

At the time of issue this report has not
been published. Any author desiring to
quote this paper should refer to it as "A
paper of the Meteorological Research
Committee (London)
M.R.P. No.819 by Mr. J.K. Bannon and
Mr. A. Gilchrist.
A copy is available in the Library
of the Meteorological Office."

M.R.P. 819
S.C.II/154
30th May, 1953

AD 37925

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

# A Statistical Study of the Variation of Temperature in the Troposphere and Lower Stratosphere

by

J.K. Bannon and A. Gilchrist (Meteorological Office)

## SUMMARY

Correlation coefficients are evaluated between temperatures at 300 mb and at three levels in the lower stratosphere (highest 60 mb) and, respectively, the pressure at the tropopause and the height of the 300 mb level, for Larkhill, Lerwick, Arctic Bay and Malta for the four mid-season months. The Dines type of relationship is confirmed for those regions and seasons affected by travelling depressions and anticyclones. Interesting seasonal changes in the stratospheric coefficients are noted at Lerwick, caused in part by the influence of the cold Arctic winter stratosphere and possibly also related to ozone amounts. Figures for the persistence of upper air temperatures are derived in the course of the analysis.

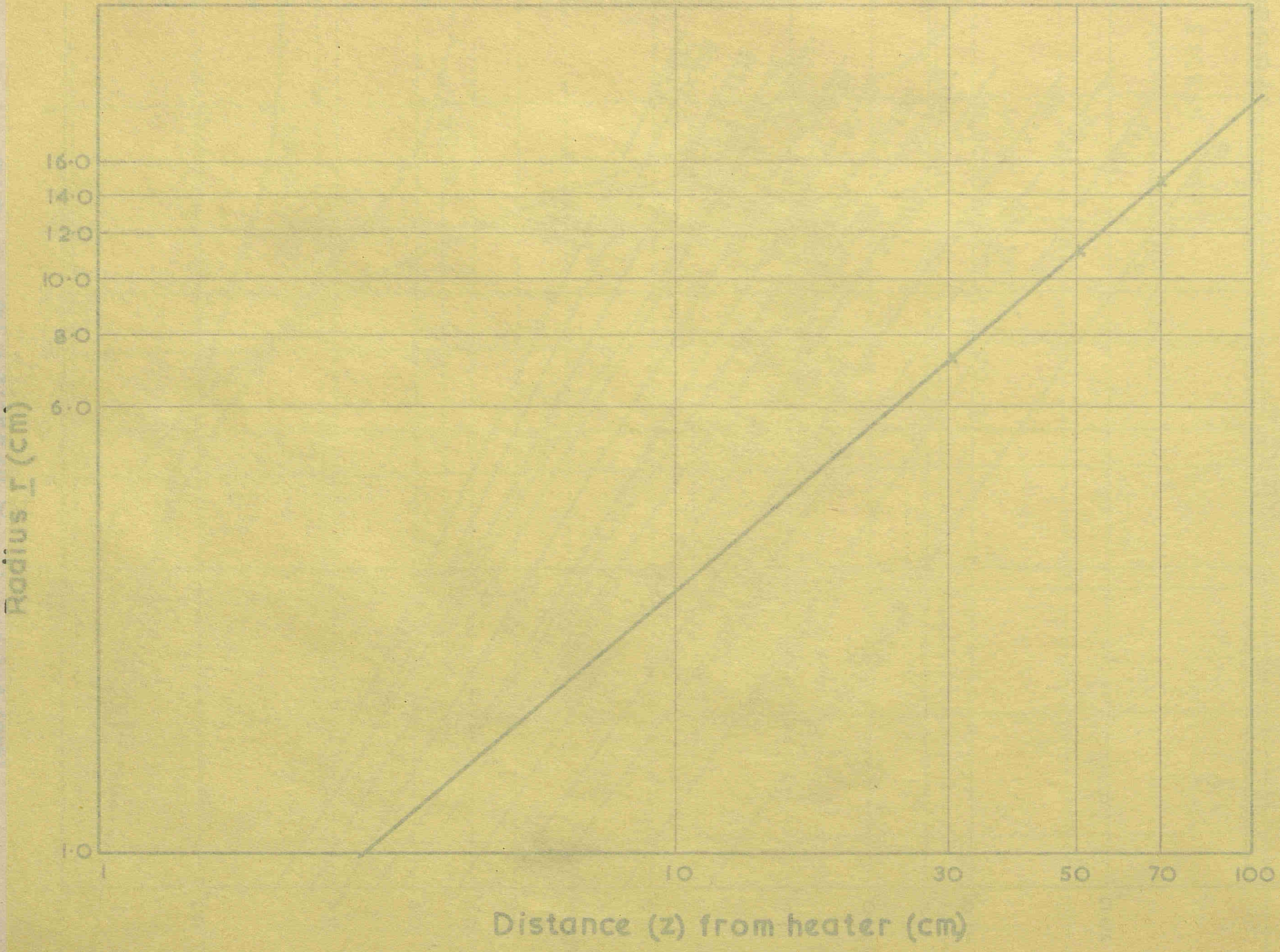
## 1. Introduction

In temperate latitudes there is a well known relationship between upper air temperatures and the general pressure distribution and level of the tropopause respectively. A cold troposphere is associated with low pressure in the upper troposphere and a low height of the tropopause, the stratosphere being correspondingly warm; a warm troposphere occurs with high pressure in the upper troposphere, a high level of the tropopause and a cold stratosphere above it. The relationship is partly demonstrated by the correlation coefficients of Dines(1) and is described in many textbooks.

Sawyer(2) has discussed the above model of upper air characteristics in connection with a classification of upper air temperature according to the pressure of the tropopause. Sawyer(3) has also discussed the relative effects of advection and vertical action in causing changes in the height of the tropopause, coming to the conclusion that they are of equal importance, other factors such as dissipation and re-formation being much less important.

Dines(1) correlation coefficients give quantitative evidence of the general model for the troposphere and his discussion gives qualitative evidence for the lower stratosphere. Ballon-sonde observations from a number of stations throughout Europe were all used together in the evaluation of the coefficients. The approximate magnitudes of these coefficients have been confirmed by Priestley(4) for southern England. Corresponding correlation coefficients have not been

FIG. 8


16·0
14·0
12·0
10·0
8·0
6·0
1·0
Radius r (cm)
1
10
30
50
70
100
Distance (z) from heater (cm)

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been published. Any author desiring to
quote this paper should refer to it as "A
paper of the Meteorological Research
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by

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## SUMMARY

Correlation coefficients are evaluated between temperatures at 500 mb and at three levels in the lower stratosphere (highest 60 mb) and, respectively, the pressure at the tropopause and the height of the 300 mb level, for Larkhill, Lerwick, Arctic Bay and Malta for the four mid-season months. The Dines type of relationship is confirmed for those regions and seasons affected by travelling depressions and anticyclones. Interesting seasonal changes in the stratospheric coefficients are noted at Lerwick, caused in part by the influence of the cold Arctic winter stratosphere and possibly also related to ozone amounts. Figures for the persistence of upper air temperatures are derived in the course of the analysis.

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Bannon(2) has discussed the above model of upper air characteristics in connection with a classification of upper air temperature according to the pressure of the tropopause. Sawyer(3) has also discussed the relative effects of advection and vertical motion in causing changes in the height of the tropopause, coming to the conclusion that they are of equal importance, other factors such as dissipation and re-formation being much less important.

Dines'(1) correlation coefficients give quantitative evidence of the general model for the troposphere and his discussion gives qualitative evidence for the lower stratosphere. Ballon-sonde observations from a number of stations throughout Europe were all used together in the evaluation of the coefficients. The approximate magnitudes of these coefficients have been confirmed by Priestley(4) for southern England. Corresponding correlation coefficients have not been evaluated for the lower stratosphere.

It has been pointed out by several writers (for example, Riehl(5)) that this model of upper air temperature variation does not always describe the temperatures at high latitudes where the stratosphere may be relatively cold above a troposphere

/which

which is also cold for its locality. It is known also that over the Antarctic in winter, temperature sometimes decreases with height up to all levels so far explored by radio-sonde and there is no stratosphere. (See, for example, Schumacher(6)). Upper air temperatures in the tropics have little variation and there is little evidence that there is much systematic variation with pressure distribution or tropopause level.

There is thus a need for quantitative test of the model of upper air temperature variation for two reasons, viz:

(i) to find for what range of latitudes the model is applicable to the atmosphere and the relative merits of the model for different latitudes and seasons.

(ii) to find how far into the stratosphere the model is applicable.

The answer to (ii) will supply, among other things, indirect information as to how high into the stratosphere the effects of the disturbances of the troposphere (depressions and anticyclones, etc.) are noticeable.

Accordingly, a series of correlation coefficients has been evaluated for four stations; Arctic Bay (73°N 85°18'W), Lerwick (60°8'N 1°11'W), Larkhill (51°11'N 1°48'W), and Malta (35°50'N 14°27'E). Observations from Aden (12°50'N 45°1'E) and a number of Pacific stations within 20° Lat. of the Equator have also been scrutinised. The correlation coefficients are between the height of the 300 mb level ($H_{300}$) and temperatures at the 500 mb level ($T_{500}$) and at three levels in the stratosphere; also between pressure at the tropopause ($P_T$) and temperature at the same levels as above. For Arctic Bay the height of the 400 mb level was substituted for the height of the 300 mb level as the tropopause is often below the 300 mb level at that station. These pairs of variables were suggested, of course, by the correlations of Dines between pressure at 9 Km. ($P_9$) and mean temperature 1-9 Km.($T_9$) (correlation coefficient = 0.95), between $P_9$ and height of the tropopause $H_c$ (0.84), between $H_c$ and $T_9$ (0.79). Priestley(4) has shown that $T_{500}$ is a good approximation to the mean temperature of the troposphere for this statistical purpose. The height of the 300 mb level is approximately equivalent to the pressure at 9 Km. and the pressure of the tropopause was used as a measure of its height.

## 2. Data

Statistics were evaluated for the four months January, April, July and October. The data used were as follows:

(i) Larkhill

Observations made four times daily (03, 09, 15, 21 h G.M.T. approximately) in the period 1948-1950 inclusive.

Statistics between $H_{300}$ and $T_{500}$, $H_{300}$ and $P_T$, and $P_T$ and $T_{500}$ were prepared from the 03h and 15h G.M.T. observations only, as the relation between these measures is well established and unlike observations at 100 and 60 mb few observations were missing from the records.

(ii) Lerwick

Observations four times daily in the period 1948-1950 inclusive.

(iii) Arctic Bay

Observations made at 03h and 15h G.M.T. daily in the period 1948-1950 inclusive. In January soundings frequently did not penetrate far into the stratosphere; for this month observations for the period 1947-1951 were used.

/iv Malta

(iv) Malta

Observations at 03h and 15h G.M.T. daily for the period 1948-1951 inclusive.

(v) Aden and tropical Pacific stations

Correlation coefficients were not attempted but observations were scrutinised for three years (1949-51) at Aden and for certain months in 1949, 1950 and 1951 for the Pacific stations.

The data for Larkhill, Lerwick, Malta and Aden are tabulated in the Upper Air Climatological Branch. Data for Arctic Bay are recorded in the Meteorological Office Library on microfilm supplied by the Canadian Meteorological Service. The observations for the Pacific islands are available in the Daily Series Synoptic Weather Maps published by the U.S. Weather Bureau.

The pressure of the tropopause is tabulated for all the observations analysed except those for the Pacific Islands; for these the tropopause was determined from the standard level readings and the temperatures recorded at "special points" when these were given. Current Meteorological Office definitions (see, for example, the Introduction to the Daily Aerological Record) could be applied without difficulty to these tropical tropopauses which are usually simple in structure.*

---

* Selecting the level of the tropopause at British or Arctic stations is usually straightforward.

---

At Malta, however, two tropopauses are commonly evident, except in summer when usually a high tropical tropopause is found at a pressure just less than 100 mb. Except in summer only the lower tropopause was normally tabulated in the period analysed; unless stated otherwise the statistics refer to this lower tropopause. The definitions used in the determination of tropopause were not sufficiently precise when the tropopause was "double" to ensure uniformity of practice and to this extent the statistics for Malta may be in error.

To assess this error the tropopauses for Malta were redetermined for the 1950,1951 data using the experimental definitions for the tropopause described in M.O.555 Part 3 (Upper Air Data: Habbaniya). The greatest number of changes in tropopause level required in any month was 28 (October 1950), but the number of changes in other months was much less, usually under 10. The effect of these changes on the statistics was not great; thus the greatest change in the mean pressure at the tropopause for a particular month was 13 mb (January 1950) and the changes in correlation coefficients between $P_T$ and $T_{500}$ and between $P_T$ and $T_{100}$ evaluated for the mid-season months of 1950, 1951 separately were no greater than was to be expected from the size of the samples worked. The largest change was from 0.42 corrected to 0.66 for October, 1950, from 48 observations having some coherence so that the number of independent observations was of the order of ten. It is concluded that the statistics for Malta for the whole period 1948-1951 are not significantly in error.

## 3. Results

The main statistical results of the investigation are given in Tables I, II, III and IV for Larkhill, Lerwick, Arctic Bay and Malta respectively. The statistics are given for each mid-season month separately. Thus in Table I the second column gives values of $\rho$, the correlation coefficient between the height of the 300 mb level ($H_{300}$) and the temperatures at 500 mb ($T_{500}$), 150 mb ($T_{150}$), 100 mb ($T_{100}$), 60 mb ($T_{60}$) and the pressure at the tropopause ($P_T$) respectively. The third column gives the mean of the values of $H_{300}$ used in the calculation of $\rho$, the fourth column gives the standard deviation ($\sigma_H$) of these values of $H_{300}$, the fifth and sixth columns give the mean ($\bar{T}$) and the standard deviation ($\sigma_T$) respectively of the temperatures ($T_{500}$, $T_{150}$ etc.) which were correlated with $H_{300}$; the seventh column gives the numbers of observations used in the calculations. The other columns are similarly defined.

Regarding the numbers of observations used in the statistics it must be remembered that temperature observations at a fixed level above one station will show some persistence from one observation to the next. Carruthers(7) quotes a "persistence length" of three days for daily temperature readings at a fixed hour in a conventional screen near the earth's surface. Lewis and McIntosh(8) find 5 as the "equivalent number of repetitions" for the departures of the Edinburgh daily mean temperature for 1950 from the corresponding 50-year average. Craddock(9) however found little correlation between the 6-hour thickness tendency at Larkhill in the layers 1,000-700 mb and 700-500 mb and the similar tendencies after 24 hours or more.

It was necessary therefore to discover the amount of coherence (or persistence) in a series of upper air temperatures and accordingly Lewis and McIntosh's(8) procedure was applied to $T_{500}$ and $T_{150}$ for Lerwick for January and July, 1950, for Malta for January and July, 1951; also to $T_{500}$ for January, 1948 and July, 1950 and to $T_{200}$ for July, 1949, for Arctic Bay. There were insufficient observations at 200 mb at Arctic Bay in any January to make an estimate of persistence. The results are given in Table V, the persistence being calculated from the Formula (2) of that paper for the "equivalent number of repetitions".

Table V

Persistence (in hours) of upper air temperature

| | January | | July | |
|---|---|---|---|---|
| | 500 mb. | 150 mb. | 500 mb. | 150 mb. |
| Lerwick<br>(1950) | 36<br>Q | 40<br>Q | 48<br>A | 27<br>Q |
| Malta<br>(1951) | 66<br>A | 42<br>A | 60<br>A | 72<br>A |
| Arctic Bay | 500 mb. | 200 mb. | 500 mb. | 200 mb. |
| | 66 (1948)<br>A | - | 66(1950)<br>A | 60(1949)<br>Q |

A = s.d. of temperature near average

Q = " " " more than 1°F. less than average.

The figures can only give a rough estimate of the persistence as presumably this will depend to some extent on the character of the weather in the particular month chosen for scrutiny. The variability of temperature in those particular months is some indication of their character and the letters A (average) and Q (quiet) are used for this purpose in the Table, A signifying that the standard deviation of temperature is roughly the same, and Q that it is more than 1°F. less than the appropriate long-term standard deviation.

It was thought that persistence should be greater in quiet settled weather. Accordingly the persistence was estimated for $T_{500}$ for Larkhill for the period 12th-31st August, 1947, when the weather was mainly anticyclonic and fine over southern England. Surprisingly the persistence was less than 24 hours, the standard deviation of temperature being about half that of the long-term value.

It would seem that the number of independent observations in the Lerwick and Larkhill data is about 45-50 (dividing the total number by seven or eight if a persistence of nearly two days is assumed and there are four observations per day). For Malta and Arctic Bay the number of independent observations is only 30-40. For $T_{100}$ and $T_{60}$ where the numbers of observations are much reduced, the series of observations will have less coherence than a full series; it is impossible to estimate the correct number of observations without investigating where observations are missing in the series, but assuming a fair measure of scatter of the observations throughout the period, the number of independent observations may be assumed to be 30-35 at 60 mb and 35-40 at 100 mb at Larkhill and Lerwick, but rather fewer at Malta. In the same way the number of independent observations at 150 mb at Arctic Bay (Table III) may be estimated as 20-25 in January and 30 in July.

Casual errors or the regular diurnal variation of the temperatures may be expected to reduce the correlation coefficients. The reduction will only be appreciable, however, when these errors or the diurnal variations are such that the scale of the ratio of their standard deviation to the standard deviation of temperature is comparable with unity. For Larkhill(10), Lerwick(11) and Malta(12) diurnal variation of temperature about the mean at 100 mb (the highest level for which there are reliable data for the last two), is less than 3°F.; assuming the standard deviation of the diurnal variation to be about 3°F., its ratio to the standard deviation of temperature (all observations) is at most 0.4 and thus negligible. For lower levels the ratio is smaller. Casual instrumental errors of the British type radiosonde have a standard deviation of 2°F. in the stratosphere and considerably less in the troposphere(13) so that their effect on the correlation coefficients will be even less than that of the diurnal variation. At 60 mb the diurnal variation at Larkhill is about 4.7°F. in July and the corresponding standard deviation of temperature is 6.5°F. so that the standard deviation of the diurnal variation is of the same order as the standard deviation of temperature when "corrected" for diurnal variation. The correlation coefficients for $T_{60}$ might therefore be reduced in the ratio $1/\sqrt{2}$ from what might be called the "true" values. For other months the reduction would be much smaller. It may be assumed that there may be a similar error in the coefficients for $T_{60}$ for Lerwick and Malta.

Diurnal variation of temperature in the upper air at Arctic Bay would be expected to be small, since observations are made at about 21 and 09h local time (03 and 15 G.M.T.). This was found to be the case. Differences between monthly mean-temperatures at 150 mb for the two hours for January, April, July, October for the years investigated being at most 2.7°F. and showing the 03 G.M.T. means less than the 15 G.M.T. means in January and April but the reverse in July and October, indicating no systematic diurnal variation. Radiation effects would be expected to be most noticeable in April and October when there is a proper night and day at Arctic Bay. For other levels below 150 mb, except near the ground, diurnal variation of temperature should also be small. Casual observational errors of the U.S.A. type radio-sonde used at Arctic Bay are not known but may be assumed to be of the same order of magnitude as the errors of the British radio-sonde.

As a check that diurnal variation has little effect on the correlation coefficients in Tables I - IV, the coefficients for $P_T$ against $T_{500}$ and $T_{100}$ for Lerwick and Malta, all seasons, and for $P_T$ against $T_{500}$ and $T_{200}$ for Arctic Bay, April and October only, were evaluated for the hours 03 and 15 G.M.T. separately. The results show no significant variation from the values given in Tables I - IV, the largest difference being 0.42 - 0.27 (Malta, $P_T$, $T_{100}$ for April) which is not significantly large considering the numbers of observations.

Statistics are not presented for Aden or for the Pacific stations. It was obvious that no significant correlation could be discovered from the data. For the Pacific stations the range of temperature at each of the levels 500, 300 and 200 mb for a particular month is about 10°C., variations from day to day usually being considerably less (2-3°C.). Such variations are probably not much greater than the observational errors to be expected. It may be that some relation will be found between upper air temperatures and, for example, the tropopause pressure in the Tropics for the organized disturbances, easterly waves, etc. which occur. Data from the tropical Pacific for the regions of the tropopause are only now becoming available in quantity and an analysis of the upper troposphere and lower stratosphere for such occasions may discover some correlations analogous to those for temperate latitudes. It is clear that there is little of the kind to be discovered in undisturbed weather in the troposphere. Aden, however, is not subject to such disturbances and the upper air over it appears to be comparatively steady in character from day to day. There is a little evidence that temperatures in the lower stratosphere in the tropics are more variable than in the upper troposphere and it may be that even in undisturbed weather (i.e. undisturbed in the lower stratosphere) some relation may be found between tropopause and stratosphere temperatures. The greater range of stratospheric temperatures may of course be merely the result of their greater liability to error. Observations in the lower stratosphere are as yet too few for a proper analysis.

## 4. Discussion

### (a) Larkhill

The correlation coefficients in Table I, Larkhill, confirm the findings of Dines (1) and Priestley (4) for the troposphere. Table VI below shows corresponding values.

Table VI

Correlation Coefficients Compared

| | | | |
|---|---|---|---|
| Dines (Europe) | $H_c\ T_m$ 0.79 | $H_c\ P_9$ 0.84 | $T_m\ P_9$ 0.95 |
| Priestley (Larkhill 1944) | $H_c\ t_{500}$ 0.79 | | |
| Larkhill (1948-1950) | $P_T\ T_{500}$ -0.75 | $P_T\ H_{300}$ -0.79 | $T_{500}\ H_{300}$ 0.90 |

Where $H_c$ = Height of the tropopause

$T_m$ = mean temperature from 1-9 km.

$P_9$ = the pressure at 9 km.

$t_{500}$ = the departure of temperature at 500 mb from the long period monthly mean

and other symbols have the meanings described in section 3.

The figures for Larkhill are the means of corresponding figures for the 4 months January, April, July, October separately.

Dines' statement that low tropopause is, in general, associated with warm, and high tropopause with cold lower stratosphere is also confirmed by the correlations. $P_T$ was very seldom less than 150 mb so that $T_{150}$ may be considered a stratospheric level.

The stratospheric coefficients are seen to be lowest in April when those for $T_{60}$ are so small as to have little or no significance. These same coefficients are largest in October and January; reference is made to these facts when discussing the Lerwick coefficients.

The decrease of the coefficients with height in the stratosphere is discussed below.

### (b) Lerwick

The tropospheric coefficients (Table II) again confirm Dines' well-known results. The stratospheric coefficients are noticeably less in January and April than corresponding values for Larkhill and as this tendency is general among all the stratospheric coefficients it is almost certainly real. The effect is most noticeable at 100 and 60 mb.

For most of the year the mean temperature in the lower stratosphere increases northwards over the northern hemisphere. In winter, however, the Arctic stratosphere is relatively cold so that the mean temperature distribution in January, at 150 mb for example, shows low temperature over the pole, a belt of higher temperatures in middle latitudes (axis about 55°N. over Britain) and cold again over the tropics. (Authority, unpublished charts in M.O.23). Deviations from the mean distribution occur, of course, so that on some occasions Lerwick may be in or to the south of the warm stratospheric belt and on others, following an excursion of Arctic air southward, it will be within, or on the edge of the Arctic cold stratosphere. This is well illustrated by the winds over Lerwick which on some occasions decrease with height in the stratosphere and on others increase with height when on the edge of the cold Arctic stratosphere (15). The temperature often decreases slowly with height in the stratosphere over Lerwick in winter as has been remarked by Miles (16), though on other occasions the normal increase with height is noted. Riehl(5) has shown that over Arctic Canada in winter the lower stratosphere may be colder or warmer than that to the south although the general weather pattern at the surface is similar.

This variability in the type of temperature distribution in the lower stratosphere in winter is most likely the cause of the low stratospheric coefficients in Table II in January. The explanation of the low stratospheric coefficients for April is more difficult. In April the mean temperatures in the lower stratosphere are higher over the Arctic than in low latitudes though the differences are not so marked as in summer and in autumn. A few occasions are observed in April of wind increasing with height above the tropopause at Lerwick indicating lower stratospheric temperatures to northward, but the number is so small that it is impossible to make the same deductions as for January. Certainly at this period of the year the Arctic stratosphere is warming up quickly or has just become warm rapidly. From the analogy of almost all other atmospheric processes it may be argued that such warming will not be at all uniform in time or space and that the reduced stratospheric coefficients in Table II for April are thus explained. That the reduction in the coefficients is most marked at 60 and 100 mb is explained if the warming of the Arctic stratosphere is affected by absorption of solar radiation in the ozone layer. It is well known that the amount of ozone in the upper atmosphere is greatest over the Arctic in spring and shows a rapid rise in high latitudes between January and March (see, for example, Normand(17)). The greater variability of stratospheric temperature at Lerwick in January and April is shown by the values of standard deviation of temperature at the various levels in Table II.

The stratospheric correlation coefficients in Table II are greatest in October when the amount of ozone in the stratosphere is near its minimum. The Larkhill coefficients are least in April and greatest in October and January and these facts lend further support to the tentative argument in the last paragraph.

### (c) Arctic Bay

The correlation coefficients for Arctic Bay(Table III) are of a different character from those for Larkhill and Lerwick. Roughly, in July and to a less extent in October, the coefficients are similar to those for the British Isles. In January and April, however, though $P_T$ and stratospheric temperatures are related in much the same way as at Lerwick and Larkhill, $H_{400}$ a measure of the pressure at the top of the troposphere, or alternatively of tropospheric mean temperature and surface pressure combined, is little related to temperatures in the lower stratosphere. The relation between $H_{400}$ and $P_T$ is, in general, less than the corresponding relation between $H_{300}$ and $P_T$ for Lerwick and Larkhill except in July.

In temperate latitudes the day to day changes of temperature are largely caused by advection and vertical motion. Thus, for example, a rise in temperature at Larkhill at 150 mb may be caused partly by the advection of warmer air and partly by the subsidence of air in the lower stratosphere; the latter is usually associated with a lowering of the tropopause also partly advective and partly caused by subsidence. These advective and subsidal processes show up through the correlation coefficients of Tables I and II discussed above for it is well-known that warm air carries with it a high tropopause and cold air a low tropopause and

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it is also recognised that the air in the extreme upper troposphere and the lower stratosphere subsides over depressions and rises over anticyclones(8). These relations have been discussed in more detail elsewhere (19, 3, 2).

At Arctic Bay in winter and spring there is seldom, if ever, a sequence of weather systems with changing air masses; shallow depressions and feeble anticyclones alternate at the surface. The upper air (500 mb) in these months is dominated by a low which is usually west of Greenland and southeast of Arctic Bay but which oscillates about this position and can have secondary lows associated with it. In summer, and also to a less extent in autumn, frontal depressions affect Arctic Bay but much less frequently than they do the British Isles; the main polar front disturbances are much further to the south. The uniform radiative properties of the vast area of snow or ice in the Arctic, especially in winter and spring, are probably responsible for this uniformity of air mass. It is not surprising, therefore, that the model of upper air temperature relations which applies to western Europe does not apply so well to the Arctic, north of Canada.

Reasonable deductions from the correlation coefficients of Table III are:-

(i) that temperatures in the lower stratosphere are influenced by vertical motion to the same extent as over the British Isles, or perhaps to a slightly greater extent ($P_T$ and $T_{250}$, $T_{200}$, $T_{150}$).

(ii) subsidence at the level of the tropopause is associated with pressure near the top of the troposphere ($H_{400}$, $P_T$ correlation). The association is not strong but is probably similar to that over the British Isles for the following reasons. Over Britain, advection plays some part in the association of $H_{300}$ and $P_T$; Sawyer(2) estimates it as similar in magnitude to the effect of vertical motion. The correlation between $H_{300}$ and subsidence at tropopause level would thus be approximately $1/\sqrt{2}$ times the $H_{300}$, $P_T$ correlation, i.e. about 0.5 or 0.6. Over Arctic Bay advection is probably much less important except in July; the $H_{400}$ $P_T$ correlations of about 0.6 in January, April, and October probably arise through the effect of subsidence, the July coefficient (0.81) being partly caused by advection as well.

(d) Malta

The coefficients in Table IV show another type of relation. In January and April the correlation coefficients are generally similar to those for Lerwick and Larkhill throughout the year; in July and October they are much smaller with the exception of those for $T_{500}$, $H_{300}$ which merely reflect the static relation between pressure and temperature and the fact that $T_{500}$ varies closely as the mean temperature of the troposphere up to the 300 mb level.

In winter and spring the Central Mediterranean is affected by travelling disturbances common to temperate latitudes. The correlations in Table IV reflect this-being only a little less than for Larkhill for the same months. In summer, however, Malta's weather is influenced by the sub-tropical high pressure belt, the tropopause is high (about 100 mb) and the temperate latitude correlations are not found. In October the summer weather is breaking and lower tropopauses are found again on many days (12); there is some correlation between $P_T$ and stratospheric temperatures but little relation between the other variables.

As stated previously the tropopause is often double at Malta and as a general rule the lower tropopause was selected for the present analysis. In January and April there is a low level tropopause (i.e. roughly in the range 150 - 300 mb) on almost every occasion; on perhaps one third of occasions there is an upper tropopause (60-120 mb) as well. The analysis took no account of these upper tropopauses so that the correlations involving $T_{150}$ and $T_{100}$ are not entirely representative of true stratospheric conditions as up to one third of the observations at these two levels were below the upper tropopause. In July a lower tropopause occurs on a few days and usually there is a single high tropopause. In October a few days occur with a single high tropopause; on others there may be a single low tropopause or a double tropopause. If two tropopauses were reported the lower was accepted for the present analysis so that in the October data there is a mixture of a few high level (tropical) with the predominating low level (temperate) tropopauses. This, however, does not fully explain the low correlations between $P_T$ and tropospheric parameters in October; the correlations between $P_T$ and $T_{500}$ and $P_T$ and $T_{100}$ were evaluated for the years 1950, 1951 separately using only those occasions in October when a low tropopause (i.e. $P_T$ greater than or equal to 150 mb) was found. The results are given in Table X. It is seen that though the relation between $P_T$ and $T_{500}$ was appreciable in 1950, in 1951 it was no more than the value given in Table IV. The stratospheric relation remains substantially unchanged.

### (e) Variation of the correlations with height in the stratosphere

It is seen from Tables I - IV that the strength of the relation between temperature in the lower stratosphere and $P_T$ and $H_{300}$, respectively, decreases with height. (The figures for July in Table IV, Malta are not applicable as the tropopause is usually above the 100 mb level). This decrease in the correlation coefficient with height is probably partly caused by the increasing diurnal variation of temperature at the greater heights (see Section 3) but the consistent decreasing tendency is almost certainly real. Thus, for example, at Larkhill in April (Table I) the coefficients between $H_{300}$ and $T_{60}$ and $P_T$ and $T_{60}$ are so small as to have no significance. Table VII gives the square of the ratio of the coefficients for $T_{60}$ to $T_{150}$ for Larkhill, Lerwick and Malta and for $T_{150}$ to $T_{250}$ for Arctic Bay as an indication of the reduction with height in the relation between stratospheric temperatures and $H_{300}$ and $P_T$ respectively. When the coefficients were not significant at the 5% level ( $\rho$ less than 0.34 approximately for the number of independent observations at 60 mb; see, for example, Fisher(14) ), the ratios have been omitted.

Table VII

Square of ratios of correlation coefficients at greater and lesser heights in the stratosphere

| | Jan.<br>Stratospheric temperature & | | Apr.<br>Stratospheric temperature & | | July<br>Stratospheric temperature & | | Oct.<br>Stratospheric temperature & | | Mean<br>Stratospheric temperature & | |
|---|---|---|---|---|---|---|---|---|---|---|
| | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ |
| Larkhill | .62 | .30 | - | - | .34 | .21 | .45 | .48 | .47 | .33 |
| Lerwick | - | - | - | - | .25 | .28 | .49 | .45 | .37 | .37 |
| Malta | .28 | .30 | - | - | - | - | - | - | - | - |
| | | (above for 60 mb and 150 mb) | | | | | | | | |
| Arctic Bay | - | .59 | - | .73 | .80 | .74 | - | - | - | .69 |
| | | (for 150 mb and 250 mb) | | | | | | | | |

The means of these ratios shown in the last two columns are almost certainly underestimates of the mean decrease with height of the relation between stratospheric temperatures and $H_{300}$ and $P_T$ respectively since small values of $\rho$ , neglected as insignificant, occurred mainly at 60 mb. The general deduction is therefore that the relations between temperature in the lower stratosphere and $H_{300}$ and $P_T$ respectively are reduced by a factor of more than 2 between 150 and 60 mb at Larkhill, Lerwick and Malta. At Arctic Bay the relation between $P_T$ and $T_{250}$ is not much greater than the relation between $P_T$ and $T_{150}$.

## 5. Variation from year to year

Over the British Isles and in many other areas also, the character of the weather in a particular month varies markedly from year to year. It was thought that such variations might have a bearing on the meaning of the correlation coefficients set out in Tables I - IV, and accordingly correlation coefficients for a selection of parameters were evaluated for each year separately. Table VIII shows the various statistics for $P_T$ against $T_{500}$ and $P_T$ against $T_{100}$ for Lerwick, Table IX, the statistics for $P_T$ against $T_{500}$ and $T_{200}$ for Arctic Bay and Table X the statistics for $P_T$ against $T_{500}$ and $T_{100}$ for Malta.

### (a) Lerwick

The figures for January, 1948 are interesting. In that month there were many instances of wind increasing with height in the stratosphere indicating a cold stratosphere to northward. The coefficient for $P_T$, $T_{100}$ is so small that there can have been little relation between these two variables and that for $P_T$ and $T_{500}$ is much smaller than for 1949 and 1950. Compare the argument in Section 5 (b). The annual variations in the other coefficients do not appear significant; such variations might be expected from samples of these small sizes from one general population.

### (b) Arctic Bay

Here again with the exception of the figures for October, 1949, the coefficients show no significant variation. In October, 1949, upper air temperatures over Arctic Bay varied little; weather was anticyclonic and "quiet" and the standard deviations of $T_{500}$ and $T_{200}$ were smaller than usual. It may be presumed that the temperature variations were so small that other factors, e.g. non adiabatic processes or errors of observation, were of comparable magnitude, thus explaining the lack of correlation.

### (c) Malta

The $P_T$, $T_{500}$ coefficients show little significant variation. The more marked variation in the coefficients for $P_T$, $T_{100}$ may well be real; the behaviour of the upper atmosphere appears to vary from year to year. Thus, in April, 1951 an upper tropopause was in evidence on 27 soundings while in 1950 it occurred on more than 50 of the 60 soundings.

## 6. Conclusions

This statistical investigation has confirmed for the British Isles the relations between upper air temperatures and the level of the tropopause and the pressure in the upper troposphere respectively, which Dines deduced quantitatively for the troposphere and qualitatively for the lower stratosphere from observations made over Europe. At Arctic Bay, however, relations of the Dines type do not occur in winter and spring though they do in summer when frontal depressions affect the area and to a limited extent in autumn also. On the other hand at Malta such relations only hold in winter and spring when temperate latitude weather systems affect the Mediterranean and not at all in summer when stable sub-tropical weather prevails; in autumn though the weather is reverting towards the type of more temperate latitudes, the relations are still weak. There appear to be no relations of the Dines type in the tropics, though the observations in the lower stratosphere are insufficient to show this with certainty.

A reasonable generalisation from these results, and one which is not surprising, is that the Dines type of model for atmospheric temperatures holds only in those regions affected by the travelling depressions and anticyclones usually associated with temperate latitudes.

The seasonal variation of the correlation coefficients at Lerwick are interesting. The low correlations in January probably arise from excursions of Arctic air southwards with a cold stratosphere. The low stratospheric correlations in April, and high correlations in October are thought to be associated with ozone amounts in the high atmosphere. Further work is necessary on both these peculiarities.

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A rapid decrease of the correlation coefficients with height in the lower stratosphere has been found except at Arctic Bay where observations are not available in any number at high altitudes. This was expected. The numerical values obtained indicate that the influence at the 60 mb level of the travelling depressions and anticyclones of middle latitudes is less than 0.5 that at 150 mb.

The variation of the correlation coefficients from year to year was small in general. This has little meaning when the coefficients are small but for the larger coefficients, shows the generality of the Dines model of temperature relation.

The correlation analysis used in this investigation can only test for a linear relation between the variables. A more refined analysis could be made without difficulty which would show up non-linear relations between pairs of variables if observations were grouped by mechanical sorting into ranges of tropopause pressure and height of the 300 mb level respectively.

The peculiarities of the cold Arctic stratosphere and possible relations between stratospheric temperatures and ozone amounts can probably best be studied by synoptic methods in the first place.

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## TABLE I.

## Statistics for Larkhill 1948-1950

| | $H_{300}$ | | | | | | $P_T$ | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $\overline{H}_{300}$ | $\sigma_H$ | $\overline{T}$ | $\sigma_T$ | No.of Obs. | $\rho$ | $\overline{P}_T$ | $\sigma_{P_T}$ | $\overline{T}$ | $\sigma_T$ | No.of Obs. |
| | | (10's of ft.) | (10's of ft.) | (°F, -ve) | (°F.) | | | (mb) | (mb) | (°F, -ve) | (°F.) | |
| | | | JANUARY | | | | | | | JANUARY | | |
| $T_{500}$ | +.86 | 2970 | 77 | 10 | 8.6 | 177 | -.79 | 237 | 52 | 10 | 8.5 | 175 |
| $T_{150}$ | -.79 | 2971 | 76 | 69 | 10.4 | 322 | +.77 | 234 | 52 | 69 | 10.7 | 321 |
| $T_{100}$ | -.73 | 2982 | 73 | 73 | 8.4 | 253 | +.67 | 227 | 47 | 73 | 8.5 | 252 |
| $T_{60}$ | -.62 | 2999 | 66 | 75 | 8.4 | 69 | +.42 | 217 | 49 | 75 | 8.2 | 68 |
| $P_T$ | -.81 | 2971 | 77 | | | 175 | | | | | | |
| | | | APRIL | | | | | | | APRIL | | |
| $T_{500}$ | +.91 | 2980 | 67 | 9 | 9.6 | 174 | -.84 | 257 | 61 | 9 | 9.4 | 169 |
| $T_{150}$ | -.75 | 2983 | 66 | 62 | 8.6 | 309 | +.75 | 252 | 57 | 62 | 8.6 | 307 |
| $T_{100}$ | -.57 | 2984 | 66 | 65 | 7.3 | 260 | +.55 | 252 | 56 | 65 | 7.3 | 259 |
| $T_{60}$ | -.26 | 3008 | 54 | 64 | 7.0 | 74 | +.22 | 244 | 59 | 64 | 7.0 | 74 |
| $P_T$ | -.89 | 2978 | 67 | | | 169 | | | | | | |
| | | | JULY | | | | | | | JULY | | |
| $T_{500}$ | +.91 | 3082 | 44 | +6 | 6.0 | 183 | -.64 | 226 | 34 | +6 | 6.0 | 182 |
| $T_{150}$ | -.74 | 3082 | 44 | 60 | 7.7 | 355 | +.70 | 223 | 33 | 60 | 7.7 | 354 |
| $T_{100}$ | -.65 | 3083 | 44 | 61 | 6.6 | 308 | +.51 | 222 | 33 | 61 | 6.6 | 307 |
| $T_{60}$ | -.43 | 3099 | 41 | 58 | 6.3 | 88 | +.32 | 216 | 31 | 58 | 6.3 | 87 |
| $P_T$ | -.67 | 3082 | 44 | | | 182 | | | | | | |
| | | | OCTOBER | | | | | | | OCTOBER | | |
| $T_{500}$ | +.91 | 3040 | 66 | +1 | 9.3 | 183 | -.72 | 220 | 51 | +1 | 9.3 | 181 |
| $T_{150}$ | -.82 | 3041 | 66 | 73 | 9.4 | 354 | +.78 | 217 | 51 | 73 | 9.3 | 352 |
| $T_{100}$ | -.74 | 3040 | 64 | 73 | 7.4 | 310 | +.65 | 217 | 49 | 73 | 7.4 | 308 |
| $T_{60}$ | -.55 | 3044 | 60 | 71 | 6.4 | 105 | +.54 | 217 | 51 | 71 | 6.4 | 105 |
| $P_T$ | -.80 | 3040 | 67 | | | 181 | | | | | | |

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$H_{300}$ = height of 300 mb surface

$P_T$ = pressure at tropopause

$T_{500}$, $T_{150}$ etc. = temperature (°F.) at 500, 150 mb etc.

$\rho$ = appropriate correlation coefficient

$\sigma_H$ = standard deviation of $H_{300}$

$\sigma_{P_T}$ = standard deviation of $P_T$

$\sigma_T$ = standard deviation of T

/TABLE II.

# TABLE II

## Statistics for Lerwick 1948-1950

|  | $H_{300}$ |  |  |  |  |  | $P_T$ |  |  |  |  |  |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
|  | $\rho$ | $\bar{H}_{300}$ | $\sigma_H$ | $\bar{T}$ | $\sigma_T$ | No.of Obs. | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}$ | $\sigma_T$ | No.of Obs. |
|  |  | (10's of ft.) | (10's of ft.) | (°F,-ve) | (°F.) |  |  | (mb) | (mb) | (°F,-ve) | (°F.) |  |
|  |  |  | JANUARY |  |  |  |  |  |  | JANUARY |  |  |
| $T_{500}$ | +.85 | 2895 | 75 | 18 | 10.3 | 361 | -.78 | 265 | 61 | 18 | 10.3 | 356 |
| $T_{150}$ | -.53 | 2897 | 75 | 68 | 9.6 | 315 | +.69 | 261 | 62 | 68 | 9.6 | 315 |
| $T_{100}$ | -.27 | 2900 | 77 | 70 | 7.7 | 245 | +.42 | 266 | 62 | 70 | 7.7 | 245 |
| $T_{60}$ | -.07 | 2922 | 73 | 73 | 7.3 | 43 | +.05 | 255 | 61 | 73 | 7.3 | 43 |
| $P_T$ | -.76 | 2896 | 75 |  |  | 356 |  |  |  |  |  |  |
|  |  |  | APRIL |  |  |  |  |  |  | APRIL |  |  |
| $T_{500}$ | +.86 | 2912 | 65 | 16 | 9.0 | 351 | -.72 | 291 | 60 | 16 | 8.9 | 343 |
| $T_{150}$ | -.46 | 2910 | 64 | 54 | 8.9 | 310 | +.64 | 291 | 60 | 56 | 8.9 | 309 |
| $T_{100}$ | -.21 | 2908 | 62 | 57 | 7.6 | 237 | +.42 | 291 | 59 | 57 | 7.6 | 236 |
| $T_{60}$ | -.17 | 2926 | 61 | 55 | 7.7 | 40 | +.18 | 286 | 63 | 55 | 7.7 | 40 |
| $P_T$ | -.74 | 2911 | 65 |  |  | 341 |  |  |  |  |  |  |
|  |  |  | JULY |  |  |  |  |  |  | JULY |  |  |
| $T_{500}$ | +.91 | 3033 | 51 | +1 | 7.1 | 367 | -.82 | 245 | 35 | +1 | 7.0 | 358 |
| $T_{150}$ | -.81 | 3032 | 51 | 51 | 7.2 | 338 | +.75 | 245 | 35 | 51 | 7.2 | 336 |
| $T_{100}$ | -.67 | 3032 | 51 | 51 | 5.3 | 298 | +.56 | 245 | 36 | 51 | 5.3 | 296 |
| $T_{60}$ | -.41 | 3046 | 60 | 49 | 5.5 | 97 | +.40 | 237 | 38 | 49 | 5.5 | 96 |
| $P_T$ | -.81 | 3029 | 51 |  |  | 358 |  |  |  |  |  |  |
|  |  |  | OCTOBER |  |  |  |  |  |  | OCTOBER |  |  |
| $T_{500}$ | +.86 | 2967 | 71 | 8 | 11.0 | 367 | -.74 | 247 | 57 | 8 | 11.1 | 362 |
| $T_{150}$ | -.83 | 2967 | 71 | 63 | 9.4 | 343 | +.78 | 247 | 57 | 63 | 9.4 | 343 |
| $T_{100}$ | -.75 | 2965 | 72 | 64 | 7.1 | 292 | +.70 | 248 | 58 | 64 | 7.1 | 293 |
| $T_{60}$ | -.58 | 2966 | 67 | 65 | 6.6 | 79 | +.52 | 243 | 56 | 65 | 6.6 | 79 |
| $P_T$ | -.82 | 2966 | 71 |  |  | 362 |  |  |  |  |  |  |

For explanation of symbols see Table I.

/TABLE III

## TABLE III

## Statistics for Arctic Bay 1948-1950

| | $H_{400}$ | | | | | | $P_T$ | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $\bar{H}_{400}$ | $\sigma_H$ | $\bar{T}$ | $\sigma_T$ | No. of Obs. | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}$ | $\sigma_T$ | No. of Obs. |
| | | (10's of ft.) | (10's of ft.) | (°F,-ve) | (°F.) | | | (mb) | (mb) | (°F,-ve) | (°F.) | |
| | | JANUARY (1947-1951) | | | | | | JANUARY (1947-1951) | | | | |
| $T_{500}$ | +.88 | 2123 | 35 | 45 | 8.1 | 260 | -.53 | 298 | 44 | 45 | 7.8 | 191 |
| $T_{250}$ | -.39 | 2126 | 40 | 73 | 11.0 | 164 | +.73 | 294 | 45 | 76 | 11.7 | 155 |
| $T_{200}$ | -.36 | 2129 | 40 | 71 | 11.1 | 96 | +.75 | 302 | 49 | 71 | 11.1 | 96 |
| $T_{150}$ | -.10 | 2136 | 39 | 71 | 11.1 | 44 | +.56 | 297 | 52 | 71 | 11.1 | 44 |
| $P_T$ | -.54 | 2121 | 38 | | | 191 | | | | | | |
| | | APRIL | | | | | | APRIL | | | | |
| $T_{500}$ | +.87 | 2175 | 57 | 40 | 9.3 | 174 | -.63 | 327 | 47 | 40 | 9.2 | 161 |
| $T_{250}$ | -.27 | 2175 | 57 | 62 | 12.6 | 150 | +.68 | 326 | 48 | 62 | 12.6 | 150 |
| $T_{200}$ | -.25 | 2175 | 59 | 55 | 11.3 | 129 | +.63 | 327 | 50 | 55 | 11.3 | 129 |
| $T_{150}$ | -.28 | 2178 | 58 | 52 | 9.1 | 113 | +.58 | 327 | 49 | 52 | 9.1 | 113 |
| $P_T$ | -.67 | 2175 | 56 | | | 161 | | | | | | |
| | | JULY | | | | | | JULY | | | | |
| $T_{500}$ | +.84 | 2332 | 41 | 9 | 7.4 | 180 | -.81 | 272 | 44 | 9 | 7.5 | 165 |
| $T_{250}$ | -.77 | 2337 | 41 | 56 | 10.6 | 166 | +.77 | 271 | 44 | 56 | 10.7 | 164 |
| $T_{200}$ | -,78 | 2333 | 41 | 49 | 9.4 | 164 | +.77 | 271 | 44 | 49 | 9.4 | 162 |
| $T_{150}$ | -.69 | 2332 | 41 | 46 | 6.2 | 153 | +.66 | 272 | 44 | 46 | 6.2 | 151 |
| $P_T$ | -.81 | 2332 | 42 | | | 165 | | | | | | |
| | | OCTOBER | | | | | | OCTOBER | | | | |
| $T_{500}$ | +.65 | 2184 | 36 | 32 | 7.7 | 132 | -.63 | 310 | 38 | 33 | 7.9 | 93 |
| $T_{250}$ | -.56 | 2185 | 36 | 60 | 8.7 | 108 | +.64 | 309 | 37 | 58 | 8.2 | 88 |
| $T_{200}$ | -.44 | 2184 | 34 | 56 | 6.5 | 90 | +.54 | 310 | 36 | 55 | 5.9 | 75 |
| $T_{150}$ | -.25 | 2187 | 33 | 57 | 5.3 | 68 | +.41 | 305 | 35 | 53 | 5.4 | 56 |
| $P_T$ | -.55 | 2183 | 36 | | | 93 | | | | | | |

$H_{400}$ = height of 400 mb surface

Other symbols as on Table I except
$\sigma_H$ = standard deviation of $H_{400}$

/TABLE IV

# TABLE IV

## Statistics for Malta 1948-1951

| | $H_{300}$ | | | | | | $P_T$ | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $\bar{H}_{300}$ | $\sigma_H$ | $\bar{T}$ | $\sigma_T$ | No. of Obs. | $\rho$ | $\bar{P}$ | $\sigma_{P_T}$ | $\bar{T}$ | $\sigma_T$ | No. of Obs. |
| | | (10's of ft.) | (10's of ft.) | ($^oF$,-ve) | ($^oF$.) | | | (mb) | (mb) | ($^oF$,-ve) | ($^oF$.) | |
| | | | JANUARY | | | | | | JANUARY | | | |
| $T_{500}$ | +.87 | 3009 | 42 | 6 | 6.3 | 197 | -.69 | 235 | 46 | 6 | 6.3 | 197 |
| $T_{150}$ | -.64 | 3007 | 42 | 68 | 7.6 | 189 | +.68 | 234 | 45 | 68 | 7.6 | 189 |
| $T_{100}$ | -.55 | 3011 | 41 | 75 | 5.9 | 162 | +.47 | 233 | 45 | 75 | 5.9 | 162 |
| $T_{60}$ | -.34 | 3021 | 37 | 78 | 6.1 | 58 | +.37 | 222 | 47 | 78 | 6.1 | 58 |
| $P_T$ | -.72 | 3005 | 42 | | | 196 | | | | | | |
| | | | APRIL | | | | | | APRIL | | | |
| $T_{500}$ | +.90 | 3052 | 38 | 0 | 5.9 | 182 | -.73 | 216 | 39 | 0 | 6.0 | 175 |
| $T_{150}$ | -.61 | 3052 | 40 | 69 | 7.6 | 171 | +.71 | 215 | 39 | 69 | 7.6 | 170 |
| $T_{100}$ | -.41 | 3051 | 37 | 74 | 5.5 | 142 | +.42 | 215 | 38 | 74 | 5.5 | 142 |
| $T_{60}$ | -.11 | 3054 | 38 | 72 | 5.5 | 43 | +.19 | 218 | 41 | 72 | 5.5 | 43 |
| $P_T$ | -.78 | 3051 | 39 | | | 175 | | | | | | |
| | | | JULY | | | | | | JULY | | | |
| $T_{500}$ | +.69 | 3167 | 23 | +18 | 4.0 | 191 | -.18 | 99 | 19 | +19 | 4.2 | 142 |
| $T_{150}$ | | 3167 | 23 | 72 | 4.8 | 182 | -.36 | 99 | 19 | 72 | 4.5 | 142 |
| $T_{100}$ | -.27 | 3167 | 23 | 85 | 5.0 | 171 | +.11 | 98 | 19 | 85 | 4.8 | 141 |
| $T_{60}$ | -.30 | 3172 | 22 | 79 | 7.4 | 97 | +.34 | 96 | 19 | 80 | 7.2 | 87 |
| $P_T$ | -.16 | 3169 | 21 | | | 142 | | | | | | |
| | | | OCTOBER | | | | | | OCTOBER | | | |
| $T_{500}$ | +.79 | 3096 | 31 | +7 | 4.8 | 216 | -.45 | 182 | 31 | +7 | 4.4 | 203 |
| $T_{150}$ | -.38 | 3096 | 31 | 76 | 6.1 | 206 | +.62 | 181 | 32 | 76 | 6.8 | 202 |
| $T_{100}$ | -.27 | 3100 | 31 | 78 | 6.3 | 178 | +.38 | 180 | 30 | 78 | 6.3 | 177 |
| $T_{60}$ | -.07 | 3108 | 26 | 76 | 6.3 | 79 | -.12 | 179 | 21 | 76 | 6.3 | 79 |
| $P_T$ | -.35 | 3096 | 31 | | | 204 | | | | | | |

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For explanation of symbols see Table I.

/TABLE VIII

# TABLE VIII

## Variation of Statistics from Year to Year

## Lerwick 1948-1950

| | $P_T$ $T_{500}$ | | | | | | $P_T$ $T_{100}$ | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Year | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}_{500}$ | $\sigma_T$ | No. of Obs. | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}_{100}$ | $\sigma_T$ | No. of Obs. |
| | (-ve ) | (mb) | (mb) | (°F,-ve) | (°F.) | | | (mb) | (mb) | (°F,-ve) | (°F.) | |
| | | JANUARY | | | | | | | JANUARY | | | |
| 1948 | 0.45 | 291 | 40 | 25 | 7.3 | 114 | 0.24 | 297 | 44 | 69 | 8.6 | 69 |
| 1949 | 0.86 | 251 | 78 | 17 | 11.6 | 122 | 0.61 | 259 | 79 | 74 | 8.7 | 78 |
| 1950 | 0.78 | 253 | 50 | 14 | 7.4 | 120 | 0.42 | 250 | 47 | 68 | 4.3 | 98 |
| | | APRIL | | | | | | | APRIL | | | |
| 1948 | 0.73 | 287 | 59 | 15 | 9.3 | 111 | 0.47 | 287 | 60 | 55 | 5.2 | 57 |
| 1949 | 0.60 | 270 | 54 | 14 | 9.1 | 116 | 0.32 | 268 | 51 | 61 | 9.4 | 83 |
| 1950 | 0.79 | 317 | 59 | 19 | 7.3 | 116 | 0.43 | 315 | 55 | 56 | 5.6 | 96 |
| | | JULY | | | | | | | JULY | | | |
| 1948 | 0.81 | 245 | 37 | 0 | 7.2 | 116 | 0.61 | 244 | 36 | 52 | 6.1 | 93 |
| 1949 | 0.83 | 234 | 35 | +4 | 6.6 | 119 | 0.51 | 233 | 36 | 52 | 5.3 | 100 |
| 1950 | 0.77 | 257 | 29 | 2 | 5.8 | 123 | 0.41 | 258 | 31 | 48 | 3.1 | 103 |
| | | OCTOBER | | | | | | | OCTOBER | | | |
| 1948 | 0.72 | 255 | 54 | 10 | 11.2 | 122 | 0.63 | 255 | 52 | 64 | 6.7 | 101 |
| 1949 | 0.76 | 236 | 59 | 6 | 10.6 | 118 | 0.79 | 236 | 60 | 65 | 7.4 | 99 |
| 1950 | 0.73 | 251 | 57 | 8 | 9.2 | 122 | 0.68 | 253 | 60 | 64 | 7.1 | 93 |

For explanation of symbols see Table I.

/TABLE IX

## TABLE IX

## Variation of Statistics from Year to Year

## Arctic Bay 1948-1950

|  | $P_T$ $T_{500}$ |  |  |  |  |  | $P_T$ $T_{200}$ |  |  |  |  |  |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Year | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}_{500}$ | $\sigma_T$ | No. of Obs. | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}_{200}$ | $\sigma_T$ | No. of Obs. |
|  | (-ve ) | (mb) | (mb) | (°F,-ve) | (°F.) |  |  | (mb) | (mb) | (°F,-ve) | (°F.) |  |
|  |  | JANUARY (1947-1951) |  |  |  |  |  | JANUARY (1947-1951) |  |  |  |  |
| 1947 | 0.67 | 275 | 29 | 39 | 5.9 | 39 | 0.61 | 278 | 29 | 79 | 7.1 | 12 |
| 1948 | 0.50 | 320 | 45 | 44 | 8.3 | 40 | 0.55 | 331 | 52 | 64 | 10.3 | 22 |
| 1949 | 0.67 | 279 | 30 | 48 | 7.2 | 37 | 0.50 |  |  |  |  | 5 |
| 1950 | 0.84 | 324 | 40 | 49 | 6.9 | 26 | 0.89 | 316 | 40 | 72 | 7.8 | 13 |
| 1951 | 0.45 | 300 | 46 | 44 | 6.6 | 49 | 0.78 | 297 | 46 | 71 | 10.7 | 44 |
|  |  | APRIL |  |  |  |  |  | APRIL |  |  |  |  |
| 1948 | 0.65 | 303 | 35 | 38 | 10.2 | 51 | 0.47 | 297 | 36 | 63 | 8.1 | 36 |
| 1949 | 0.66 | 326 | 39 | 42 | 7.0 | 56 | 0.58 | 327 | 41 | 58 | 9.7 | 45 |
| 1950 | 0.76 | 350 | 53 | 39 | 9.8 | 54 | 0.71 | 350 | 55 | 45 | 5.6 | 48 |
|  |  | JULY |  |  |  |  |  | JULY |  |  |  |  |
| 1948 | 0.76 | 251 | 37 | 6 | 6.7 | 57 | 0.70 | 251 | 38 | 54 | 9.1 | 56 |
| 1949 | 0.67 | 278 | 36 | 11 | 6.4 | 59 | 0.67 | 277 | 36 | 48 | 6.1 | 58 |
| 1950 | 0.90 | 288 | 50 | 11 | 8.3 | 49 | 0.84 | 288 | 51 | 43 | 8.9 | 48 |
|  |  | OCTOBER |  |  |  |  |  | OCTOBER |  |  |  |  |
| 1948 | 0.59 | 296 | 28 | 31 | 6.0 | 24 | 0.57 | 302 | 22 | 57 | 7.0 | 21 |
| 1949 | 0.18 | 301 | 23 | 29 | 3.2 | 24 | 0.13 | 302 | 23 | 59 | 4.6 | 20 |
| 1950 | 0.66 | 322 | 45 | 34 | 10.0 | 45 | 0.64 | 321 | 45 | 53 | 5.4 | 34 |

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For explanation of symbols see Table I.

/TABLE X

# TABLE X

## Variation of Statistics from Year to Year

## Malta 1948-1951

| | $P_T$ $T_{500}$ | | | | | | $P_T$ $T_{100}$ | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Year | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}_{500}$ | $\sigma_T$ | No. of Obs. | $\rho$ | $\bar{P}_T$ | $\sigma_{P_T}$ | $\bar{T}_{100}$ | $\sigma_T$ | No. of Obs. |
| | (-ve ) | (mb) | (mb) | (°F,-ve) | (°F.) | | | (mb) | (mb) | (°F,-ve) | (°F.) | |
| | | JANUARY | | | | | | | | JANUARY | | |
| 1948 | 0.62 | 223 | 45 | 3 | 7.0 | 47 | 0.54 | 220 | 44 | 75 | 5.6 | 38 |
| 1949 | 0.71 | 245 | 49 | 5 | 5.9 | 34 | 0.28 | 236 | 46 | 73 | 5.1 | 22 |
| 1950 | 0.70 | 224 | 40 | 8 | 5.5 | 50 | 0.27 | 223 | 42 | 76 | 5.1 | 46 |
| 1951 | 0.72 | 239 | 40 | 7 | 5.9 | 62 | 0.65 | 238 | 40 | 75 | 6.6 | 56 |
| | | APRIL | | | | | | | | APRIL | | |
| 1948 | 0.78 | 221 | 44 | +1 | 4.7 | 26 | 0.53 | 224 | 48 | 69 | 4.1 | 20 |
| 1949 | 0.61 | 208 | 28 | +2 | 5.5 | 44 | 0.14 | 205 | 26 | 76 | 5.1 | 38 |
| 1950 | 0.85 | 239 | 46 | 2 | 6.7 | 47 | 0.21 | 236 | 29 | 72 | 5.1 | 38 |
| 1951 | 0.72 | 207 | 26 | 0 | 5.6 | 56 | 0.44 | 208 | 26 | 75 | 4.2 | 44 |
| | | JULY | | | | | | | | JULY | | |
| 1948 | +0.13 | 112 | 24 | +20 | 3.3 | 12 | | | | | | |
| 1949 | 0.18 | 117 | 18 | +18 | 4.8 | 30 | | | | | | |
| 1950 | 0.22 | 093 | 14 | +20 | 3.5 | 47 | | | | | | |
| 1951 | 0.09 | 101 | 11 | +17 | 3.5 | 54 | | | | | | |
| | | OCTOBER | | | | | | | | OCTOBER | | |
| 1948 | 0.34 | 182 | 27 | +9 | 4.4 | 46 | 0.29 | 178 | 24 | 80 | 7.2 | 38 |
| 1949 | 0.45 | 200 | 29 | +5 | 3.1 | 50 | 0.43 | 197 | 28 | 77 | 4.7 | 45 |
| 1950 | 0.66 | 177 | 31 | +9 | 4.9 | 48 | 0.42 | 178 | 31 | 78 | 7.1 | 43 |
| 1951 | 0.36 | 181 | 27 | +6 | 4.1 | 50 | 0.40 | 178 | 23 | 78 | 5.7 | 41 |

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For explanation of symbols see Table I.

$$P_n = P_{n_s} + n(D)$$

$$\underline{dP} = \underline{dP_s} + \underline{dn(D)}$$

|  |  |  |  |  |
|---|---|---|---|---|
| 18000 | 17500 | +500 | T H | n |

At the time of issue this report has not been published. Any author desiring to [illegible]
[illegible]
[illegible]
[illegible] the Director
of the Meteorological Office".

M.R.P. 860
S.C.II/167
[illegible] March, 1954

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

# Variation of temperature in the lower stratosphere above Larkhill and Lerwick

(Addendum to M.R.P. 819)

by

J.K. Bannon (Meteorological Office)

## Introduction and Summary

The numbers of observations for 60 mb. used in the analysis described in the previous paper, M.R.P. 819(1), were not large. Moreover, data for all hours of observation, day and night, were used. Because of the comparatively scanty data and the possibility that radiation errors might result in correlation coefficients being erroneously small, the stratospheric correlation coefficients were computed again for Larkhill and Lerwick using all available data for the level of 60 mb. for specified hours of observation, day and night separately. The important result obtained previously, namely that the relation between temperature and the height of the 300 mb. surface and tropopause pressure respectively, falls off rapidly between 150 and 60 mb. was confirmed, but the seasonal variations in these relations appear to be less marked than was indicated in the earlier paper.

## 1. Data

As in M.R.P. 819 the analysis was restricted to data for the four mid-season months, January, April, July and October. All observations reaching the level of 60 mb. at 00 and 12h. G.M.T. before 1948 and at 03 and 15h. G.M.T. in 1948 and later years at Larkhill and Lerwick were used and values of the height of the 300 mb. surface ($H_{300}$), the pressure of the tropopause ($P_T$), and temperatures at levels 150, 100 and 60 mb. ($T_{150}$, $T_{100}$, $T_{60}$ respectively) were extracted, the day and night observations being treated separately. For January and April at Lerwick and April at Larkhill so few observations were available for the selected hours that it was necessary to use data for other hours of observation, the data being grouped according to whether the observations were in daylight or darkness. These data provided comparable sets of observations at each of the three levels, 150, 100 and 60 mb. in contrast with the data used in M.R.P. 819 which embraced many more observations at 150 mb. than at 60 mb. [illegible] there was always the possibility that observations at the higher levels, through some bias or another, might not be in the same class as those at lower levels.

$\underline{\Delta P} = \underline{\Delta P_s} + \underline{\Delta p_m(t)}$

18000 | 17500 | +500 | 7 4 | [illegible]

B/7

M.R.P. 820
S.C.III/156
10th June, 1953.

~~AD~~

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No........... by Mr.............................. A copy is available in the Library of the Meteorological Office."

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## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# Five Flights through a Thunderstorm Belt

by

R.F. Jones (Meteorological Office)

## 1. Introduction

In a previous report (1) on cumulonimbus investigation the writer gave the results of some integrations of the accelerometer records obtained in a Spitfire aircraft on flights through cumulus and cumulonimbus clouds. The curves obtained, curves of vertical speed, required adjustment to take account of any errors in reading the accelerometer trace and it was necessary to make several plausible assumptions, as to the changes of height of the aircraft, to obtain reasonable curves of the vertical speeds to which the aircraft was subjected. Lack of information of height changes experienced precluded any check on the validity of these assumptions.

With the change of aircraft type for these investigations from Spitfire to Meteor VII it became possible to carry an observer and to install an auto-recorder to give photographically at 10 second intervals the readings of I.C.A.N. height, air temperature and indicated air speed. Any accelerometer record from the Meteor, therefore, can be integrated and the curve obtained adjusted to ensure that the vertical speeds obtained give, by a further integration, correct height changes over successive periods of 10 seconds or more.

By performing a few integrations of accelerometer records obtained in the Meteor it is therefore possible to check whether the assumptions made in the previous report were reasonable as well as obtain information on vertical currents which will be of interest on their own account.

Five flights made on 13th June, 1952 through a belt of thunderstorms were chosen for analysis in this way and are the subject of this report.

## 2. Synoptic situation and flight programme

The whole of England and Wales was covered by a shallow depression forming part of a low pressure belt extending from Scandinavia to the Azores. The upper winds were variable and generally light up to 12,000 ft. while the 500 mb. and 300 mb. contour charts showed the existence of a cold pool over SW. England and the western English Channel. Conditions were therefore favourable for outbreaks of thundery rain or thunderstorms and these were confirmed by the extensive radar weather echoes visible at East Hill. Later in the day the echoes became less extensive and it was decided to investigate a belt of radar echo lying NW. to SE. to the northeast of East Hill (see P.P.I. photograph taken at 1607 GMT). The radar echo tops (see height-range photographs) reached to 35,000 to 36,000 ft., which was probably the tropopause height (1500 GMT tropopause heights: Larkhill 32,300 ft., Hemsby 39,000 ft.).

The aircraft was in the vicinity of the clouds giving this belt of echo at 1525 GMT and described them as a line of five cumulonimbus clouds lying NW. to SE. There was extensive anvil cirrus.

The first two runs were made at 25,200 ft. through the last but one towards the SE. of the cumulonimbus clouds on headings of 270° and 090° but, because the aircraft came too close to the minimum radar visible range, the three subsequent runs were made through the last cumulonimbus of all on headings of 260°, 080°, 260° at 20,200, 20,200 and 18,200 ft. respectively. All the runs took place between 1528 and 1604 GMT and the tracks followed by the aircraft through the echoing volume, after adjustment for echo movement, are shown in the P.P.I. photograph. From the appearance of the radar echo it is certain that the clouds to SE. of the belt were the most recently developed (on the P.P.I. photograph reproduced, taken just after the runs were completed, yet another new cell is seen to be developing at the southeastern extremity of the belt).

## 3. Information available

For each of the five flights analysed the following information is available.

(i) A continuous record, on 35 mm film, of accelerations in the vertical experienced by the aircraft. The speed of the film was 2.53 ft./min. (0.506 in./sec.) and the scale such that a vertical movement of the trace through 2.7 mm corresponded to an acceleration of 1g (32 ft./sec.$^2$).

The film was examined under an enlarger and the accelerations recorded to the nearest .05g as often as necessary to give a faithful reproduction down to 0.1 sec. This detail with respect to time cannot be shown accurately in Figs. I to V but the accurate time values have been used in the integrations.

(ii) The altimeter reading at 10 sec. intervals.

(iii) The indicated air speed at 10 sec. intervals. This affords an indication of the climb and dive attitude of the aircraft since the power setting of the aircraft was not usually adjusted during a run. Although constant attitude in traversing the clouds was aimed at there were nevertheless some variations in indicated air speed. Once a power setting has been chosen an indicated air speed greater than that necessary to maintain level flight indicates a diving attitude if the power setting remains unchanged. It is possible therefore to see, at least qualitatively, whether the up or down movements of the aircraft were ascribable to air currents or to changes in climb and dive attitude of the aircraft. It is hoped later to make test runs with the same aircraft to determine the variation of height with indicated air speed at a constant power setting.

(iv) The air temperature at 10 sec. intervals. The thermometer used was a ratiometer-type temperature indicator - Sangamo-Weston Type S 110 Form 9. The temperature element is a platinum wire wound on a conical head and enclosed in hydrogen. The speed corrections have been carefully determined and it is considered that the readings outside cloud are correct to ±¼°C with a lag of about 1½ sec. Measurements in cloud are, of course, subject to all the usual troubles and uncertainties (see, for example, Ref. (2)). The corrected temperature readings have also been further corrected to minimise the effect of altitude changes. This consists of adjusting the temperature readings to those that would have been experienced at the mean level of flight, by applying a dry adiabatic correction (3°C/1000 ft.) for readings outside cloud and the wet adiabatic, appropriate to the height and mean temperature of the flight, when inside cloud. As the departures from the mean level of flight were always less than 300 ft. the corrections applied on this account were never as great as 1°C.

The application of the normal speed corrections to temperatures taken within cloud means that the corrected temperatures may often be too cold by variable amounts which might be as great as 3.1°C at 18,000 ft., 2.3°C at 20,000 ft. and 0.8°C at 25,000 ft.

(v) A tape recording of the comments made by the observer in the aircraft during a cloud traverse and received from the aircraft on R/T. This includes notes of times of switching on and off accelerometer and times of entry into and exit from cloud.

(vi) Radar cross-sections in the vertical of the storm echo along the line of flight of the aircraft, taken on wavelengths of 10 cm. and 3 cm. (beam widths to half power 1.5° and 0.7° respectively in the vertical plane and 7.5° and 2° respectively in the horizontal plane). An average of eight photographs is available for each run. On many of these the aircraft echo is also visible, thus enabling the position of the aircraft relative to the radar weather echo to be known accurately at any instant.

## 4. Presentation of the data

The data for each of the five flights are shown separately in Figs. I to V respectively.

In each figure the first graph shows the accelerations experienced by the aircraft against time. The second graph is the integration of the acceleration curve giving a curve of vertical speed against time. The third graph shows the altimeter readings at 10 sec. (sometimes 9 sec.) intervals. The altimeter changes in a given time should be related to the integral of the second graph over the same period of time and the dotted lines over different time periods, usually 20 sec., on the second graph are the zero corrections which must be applied to give the correct altimeter changes. The fourth graph shows the curve of the second graph adjusted to make the altimeter changes correct. The transition from one zero correction to the next has been smoothed, taking account of the form of the original integrated curve. It is considered that this curve gives a true representation of the upward and downward speeds of movement of the aircraft. The fifth and sixth graphs of each figure give the indicated air speed and corrected air temperatures at 10 sec. (sometimes 9 sec.) intervals respectively.

Also indicated are the parts of the records obtained within cloud and the instants of entry into and departure from the echoing volume associated with the cloud. On Figs. II and V graph (4) are indicated the places where lightning was reported by the observer in the aircraft.

/5.

## 5. Evaluation of up and down air currents

The interest in the integrated curves lies in the deductions which can be made from the behaviour of the aircraft about the air currents encountered by the aircraft. If an aircraft flying level enters an upward air current and maintains its attitude it will ultimately take up the upward speed of the air, if the air current has a large enough horizontal extent. The time taken for the aircraft to take up this upward speed will depend on the characteristics of the aircraft but for the aircraft used is of the order of one second. Provided the aircraft was flying level therefore (i.e. in these cases, if the power setting was constant, provided the indicated air speed was constant) it can be inferred that the upward air speeds were at least as great as the upward speeds of the aircraft, but not much greater if the up-current took more than a second or two to cross.

Unfortunately on many occasions, presumably because of the turbulence encountered, the indicated air speed did not remain constant. On these occasions, however, the form of the integrated curve and the knowledge that increased air speed usually means diving attitude (although a change of power setting would also change the indicated air speed) enable useful deductions still to be made from the curve of aircraft vertical speed against time.

## 6. Discussion of results

### (a) Turbulence

The most striking feature of Figs. I to V is the marked change in frequency and intensity of gusts on entering cloud. In the clear air before entering and after leaving cloud there was only one gust as great as 0.4g in the five runs (at 257 sec. on Run II - see Fig. II) and this was very probably occasioned by an abrupt change in aircraft attitude (see indicated air speed trace). The frequency of detected gusts in the clear air was on the average one per 5.3 sec. (about 2,900 ft.) compared with an average frequency of gusts in cloud of one per 1.4 sec. (about 770 ft.).

All the gusts as great as 0.5g occurred within the part of the cloud from which an echo was received; the frequency of gusts in cloud but not in the echoing volume was one per 1.9 sec., i.e., gusts were slightly less frequent outside the echoing volume, but great significance cannot be attached to the result since only 66 sec. of all the flights qualified in the determination of this frequency.

### (b) Up and down air currents

The authors of "The Thunderstorm" (3) draw a clear distinction (pp. 39, 43) between gusts and "drafts". They regard the gusts as caused by eddies superimposed on the general airflow and carried along by it. On this interpretation the vertical movements of the aircraft are regarded as caused by the large scale air currents and the strength of the air current is deduced from the change of altimeter reading, while the gusts recorded are regarded as effects entirely separate with no permanent effect on the upward and downward movement of the aircraft. While it is very likely that such gusts or eddies do exist the fact remains that recorded gusts are recorded accelerations of the aircraft in the vertical and that if due account is taken of their magnitude, shape and duration, their effect on the vertical movement of the aircraft can be determined. In fact the integrated effect of all the gusts should give the speed of movement of the aircraft in the vertical. The only exception to this can be if the aircraft can enter an up or down current without recording a detectable acceleration. Mean up currents of 30 - 40 ft./sec., as determined from the aircraft altimeter trace, were quite commonly encountered in Florida and Ohio (3) and to enter these without recording a detectable acceleration, i.e. without recording a "gust", would require a very gradual increase in speed of up current with horizontal distance. For, if the minimum detectable acceleration is .05g (1.6 ft./sec.$^{-2}$), it will take more than 40 sec. for an aircraft to acquire an upward speed of 64 ft./sec., and a further 40 sec. or more for the aircraft to lose this speed, to give a mean upward speed of 32 ft./sec. without recording "gusts". With the aircraft travelling about 300 ft./sec. this would require all up or down currents of mean value 32 ft./sec. to be at least 24,000 ft. across - a very unlikely contingency (3). On the other hand two up gusts of 1.0g, each lasting for only one second and giving a mean acceleration of 0.5g for that second, would impart to the aircraft an upward speed of 32 ft./sec. Many of the recorded gusts therefore are probably not caused by small eddies superimposed on a major up or down current but are really a demonstration of the strong gradient along the line of flight of the vertical speed of the major current itself.

It seems most likely therefore that the integrated accelerometer trace (i.e. the integrated gusts) will give all the major up or down speeds of the aircraft, however caused, as well as small scale eddy fluctuations. A convincing demonstration of this given by curves (2) and (3) of Fig. II in which the three major up and down currents (after 120 sec.) shown by the integrated accelerometer record - even before corrections are applied - have their counterpart in the variation of aircraft height as shown by the altimeter record.

The outstanding feature of the curves of vertical speed is the close proximity of up and down currents of comparable magnitude. An extreme example occurs at 200 sec. on Fig. I when the aircraft changed from a maximum upward speed of 30 ft./sec. to a maximum downward speed of 23 ft./sec. in four seconds, a distance of about 750 yd., but even in the wider up currents of Fig. II, for example, there appears to be direct transition from up current to down current. A similar feature is to be observed in the integrated traces reported in M.R.P. 593 (1). Some of these transitions may be brought about by large eddies rather than by up or down currents persisting through a great depth of cloud and, while it is unlikely that all are so caused, there is no means of separating large eddies from extensive up or down currents. However they are caused, one is forced to the conclusion that within a mature cumulonimbus cloud up and down air currents as strong as 20 ft./sec. or more may exist in close proximity to each other. The danger to a pilot attempting to fly constant altitude is obvious; for to maintain altitude in an up current of 30 ft./sec., say, he must take up a diving attitude sufficient to give a rate of descent in still air of 30 ft./sec. A quick transition to a down current of 30 ft./sec. will find him descending at 60 ft./sec. and requiring a rate of climb of 60 ft./sec. to maintain altitude. Thus the effects of up and down currents and the strain on the aircraft are accentuated.

Examination of Figs. I to V shows that most of the up currents on these occasions were about 3000 to 5000 ft. across (although two were about 10,000 ft. on Run 2) and the down currents usually 1500 - 4000 ft. across. This is broadly in agreement with the experience reported in "The Thunderstorm" (3) from a much greater number of occasions. It is not therefore surprising, with the up and down currents so narrow, that flights only a few minutes later at the same height on the reciprocal track are not similar in detail of up and down currents or turbulence encountered (e.g. Figs. I and II, Figs. III and IV).

Outside cloud up and down currents would appear to have been of more variable, and usually greater, width. After allowing for variations in indicated air speed it would appear that down currents in excess of 10 ft./sec. may exist on some occasions outside cloud but up currents as great as 10 ft./sec. were only encountered within the cloud.

(c) Temperature

Detailed temperature traces exhibit a somewhat disturbing feature because of the close correlation between the curves for indicated air speed and temperature, increasing air speed being associated with colder temperatures. Since increased air speed means a greater correction to be subtracted from indicated temperature this correlation suggests that it may be a mistake to correct in detail for air speed variations, i.e., it suggests that some of the air speed readings are sudden changes, brought about perhaps by gusts, and not applicable to a time $1\frac{1}{2}$ sec. previously - the lag of the thermometer is about $1\frac{1}{2}$ sec. A variation of 10 m.p.h. in indicated air speed gives a variation of about 0.7°C in temperature correction to be applied. It is, of course, possible that the correlation between the traces is due to an external cause affecting each e.g. a cold down current would have the same effect if it caused the aircraft to take up a diving attitude.

Detailed correlation of temperature with up and down currents, even outside cloud, therefore is open to objection. Inside cloud there is the additional difficulty of knowing what speed correction to make, or even, because of the variations of water content, whether the same correction should apply in all parts of the cloud.

The mean temperatures and standard deviations for each height are shown in Table I.

Table I.

| Height (ft. ICAN) | Mean Temp. in cloud (°C) | Standard Deviation (°C) | No. of obs. | Mean Temp. clear air (°C) | Standard Deviation (°C) | No. of obs. |
|---|---|---|---|---|---|---|
| 25,200 | -30.7 | 0.6 | 27 | -30.7 | 0.6 | 31 |
| 20,200 | -19.5 | 0.9 | 12 | -19.2 | 0.7 | 23 |
| 18,200 | -16.2 | 0.7 | 7 | -15.2 | 0.3 | 12 |

As mentioned in para. 3 (iv) above the in-cloud temperatures may be too cold by up to 3.1°C, 2.3°C and 0.8°C at 18,200, 20,200 and 25,200 ft. respectively. Taking this into account it would seem likely that the mean in-cloud temperatures were warmer than the clear air at the same height with the differences in temperature not likely to exceed 2.0°C at 18,200 and 20,200 ft. and 0.8°C at 25,200 ft. Applying the full possible corrections the lapse rates in cloud would be 2.1°C/1000 ft. between 18,200 and 20,200 ft. and 2.5°C/1000 ft. between 20,200 and 25,200 ft., which are very close to the saturated adiabatic lapse rates (saturation with respect to water) for a wet bulb potential temperature of about 62°F (2.2°C, 2.4°C and 2.8°C per 1000 ft. at 18,200, 20,200 and 25,200 ft. respectively). The lapse rates in clear air, 2.0°C/1000 ft. and 2.3°C/1000 ft. between 18,200 and 20,200 ft. and 20,200 and 25,200 ft. respectively are less than the saturated adiabatic lapse rate.

It is of interest to note that the variability of temperature within cloud, as evidenced by the standard deviation of the observations, is not significantly greater than outside cloud at 20,200 and 25,200 ft.

(d) Lightning

Lightning was reported by the observer in the aircraft over the radio telephone at the time of its occurrence on one occasion on run (2) at 25,000 ft. and four occasions on run (5) at 18,200 ft. A time lag of two seconds has been allowed for the time to report and the points of occurrence are indicated in the appropriate places on curve (4) of Figs. II and V.

It will be seen that on all occasions but one (at 128 sec. on run (5)) the aircraft was not experiencing any appreciable up or down movement while on the other occasion, although it was descending at about 20 ft./sec., it seems likely, from the air speed trace, that a large part of the downward movement was occasioned by a change in attitude of the aircraft.

The fact that lightning was not reported on runs (3) and (4) which were only 2,000 ft. higher than run (5) and 11 min. and 6 min. earlier respectively suggests that run (5) was made at almost exactly the time when lightning first occurred within the cloud. From the radar echo photographs it is clear that the cloud had already reached its maximum vertical extent by the time run (3) was made.

### (e) Correlation with radar echo

Previous analyses of cumulonimbus flights have, of necessity, relied on the 10 cm. wavelength radar to determine correlations between radar echo and turbulence. The limitations of this radar, particularly in respect of the horizontal beam width of the height-range equipment, have been discussed (1). During the flights here described a 3 cm. wavelength height-range radar was also used which gave greater precision because of its narrower beam width (0.7° in the vertical and 2° in the horizontal plane). Photographs of the display of this equipment taken on each run close to the time of entry into and departure from cloud (other photographs are also available) are shown while the P.P.I. photograph gives the 10 cm. wavelength picture. It will be seen that even the small change of bearing necessary to keep the aircraft in the radar beam can substantially change the radar picture. This makes correlation with radar echo more difficult but on each photograph are indicated the areas where up or down currents of greater speed than 20 ft./sec. were encountered and where gusts of 0.5g or greater occurred. The appropriate time scale for the run is also given.

It will be seen that all gusts of 0.5g or greater were encountered within portions of the cloud giving rise to the radar echo and were often, as reported in earlier papers (1), (4), associated with discontinuities in the echo structure. There is some evidence that up currents and echo columns may coincide while on run (5) a substantial down current would seem to have existed outside the edge of the echoing volume but still within cloud; (a strong down current indicated just outside echo and cloud in run (2) would seem to have been principally caused by a change of aircraft attitude).

On runs (1) and (3) the echoing volume began about 12 sec. (about 1¼ mi.) before the observer reported entering cloud, suggesting the existence of precipitation from overhanging cloud. On runs (4) and (5) the edge of the echoing volume and cloud edge coincided while on run (2) the aircraft was in cloud about 12 sec. (about 1¼ mi.) before entering the echoing volume. On the other hand the aircraft always left the echoing volume before leaving cloud by times varying between 7 and 17 sec. (¾ - 1¾ mi.).

## 7. Conclusions and Summary

(i) It is possible to obtain reliably from an accelerometer trace and altimeter record the detailed movements of an aircraft in the vertical and thence, if indicated air speed is also recorded, deduce the strengths of up and down air currents.

(ii) The adjustments made to the integrated curves in M.R.P. 593 (1) were reasonable and the deductions made from them valid.

(iii) Up and down air currents of comparable magnitude, which may exceed 20 ft./sec. on occasions, occur close to each other in mature cumulonimbus clouds.

(iv) Down currents in excess of 10 ft./sec. may occur outside cloud but the up currents in excess of 10 ft./sec. recorded were all within cloud.

(v) There was a striking change in frequency and intensity of gusts on entry into cloud and the biggest gusts all occurred within the echoing volume of the clouds.

(vi) Mean in-cloud temperatures, although difficult to measure with certainty, were probably up to 2°C warmer than the mean environment temperature at the same height at 18,200 and 20,200 ft. and up to 0.8°C warmer at 25,200 ft.

(vii) Lightning occurred once at 25,200 ft. and four times at 18,200 ft. and on these occasions the aircraft was probably not in any appreciable up or down current. The cloud had already attained its maximum vertical development before lightning was reported.

(viii) Detailed correlation with radar echo is made difficult by the substantial changes in echo structure accompanying a small change in the azimuth of viewing when using a narrow beam width 3 cm. wavelength equipment.

## 8. References

(1) Jones, R.F.: The relation between the radar echoes from cumulus and cumulonimbus clouds and the turbulence within those clouds. Part II. M.R.P. No.593, Air Ministry, Meteorological Research Committee, London, 1950.

(2) Meteorological Office: Meteorological Air Observer's Handbook. M.O. 470, Air Ministry, London, 1945. pp. 17 - 18.

(3) Thunderstorm Project: "The Thunderstorm ". U.S. Department of Commerce, Weather Bureau, Washington, D.C., 1949.

(4) Jones, R.F.: The relation between the radar echoes from cumulus and cumulonimbus clouds and the turbulence within those clouds. M.R.P. No. 484, Air Ministry, Meteorological Research Committee, London, 1949.

M.R.P. 820.

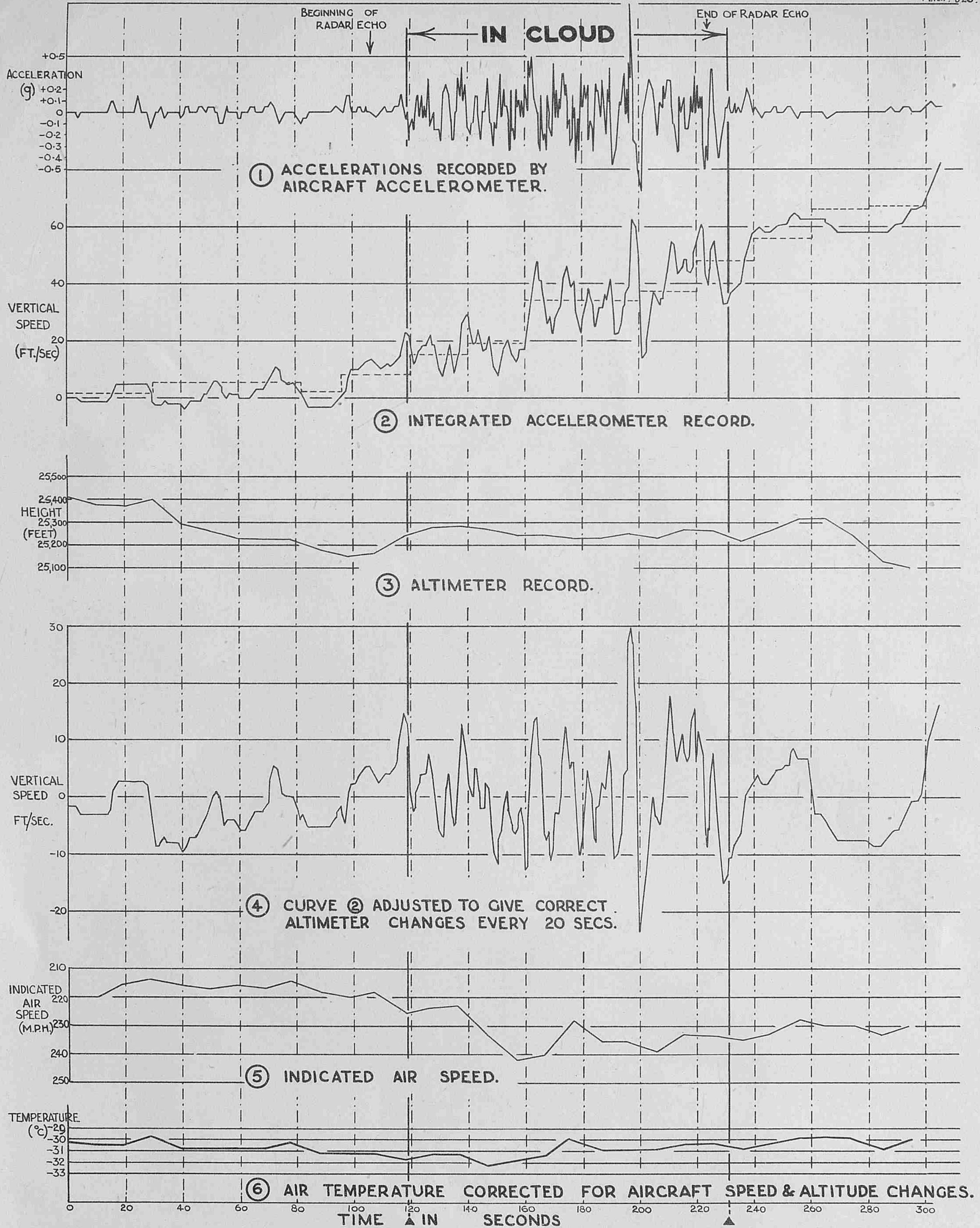


FIG.I. CUMULONIMBUS INVESTIGATION - 13TH JUNE 1952 - RUN 1. 25,200'.

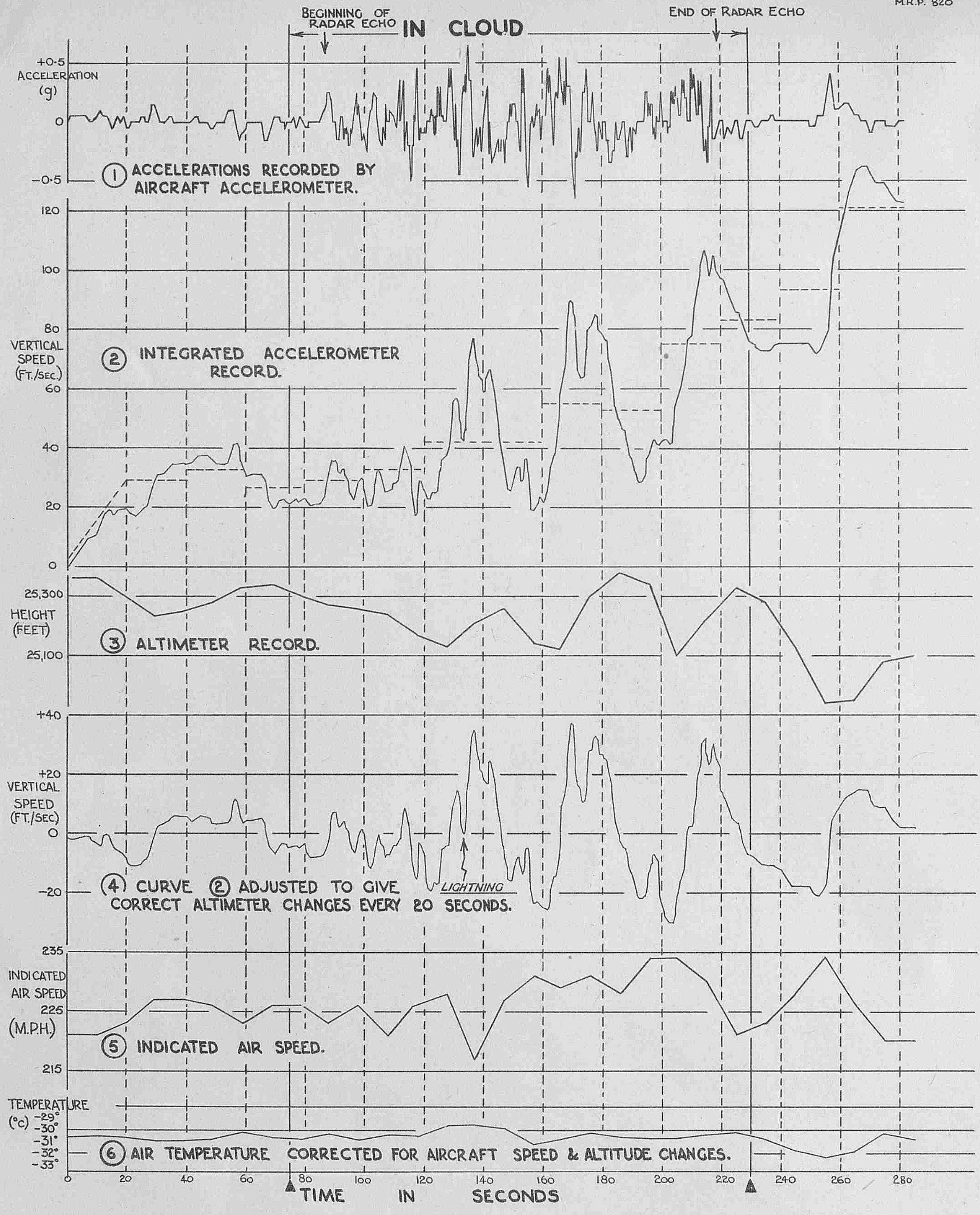


FIG. II. CUMULONIMBUS INVESTIGATION - 13TH JUNE 1952 - RUN II. 25,200'.

M.R.P. 820.

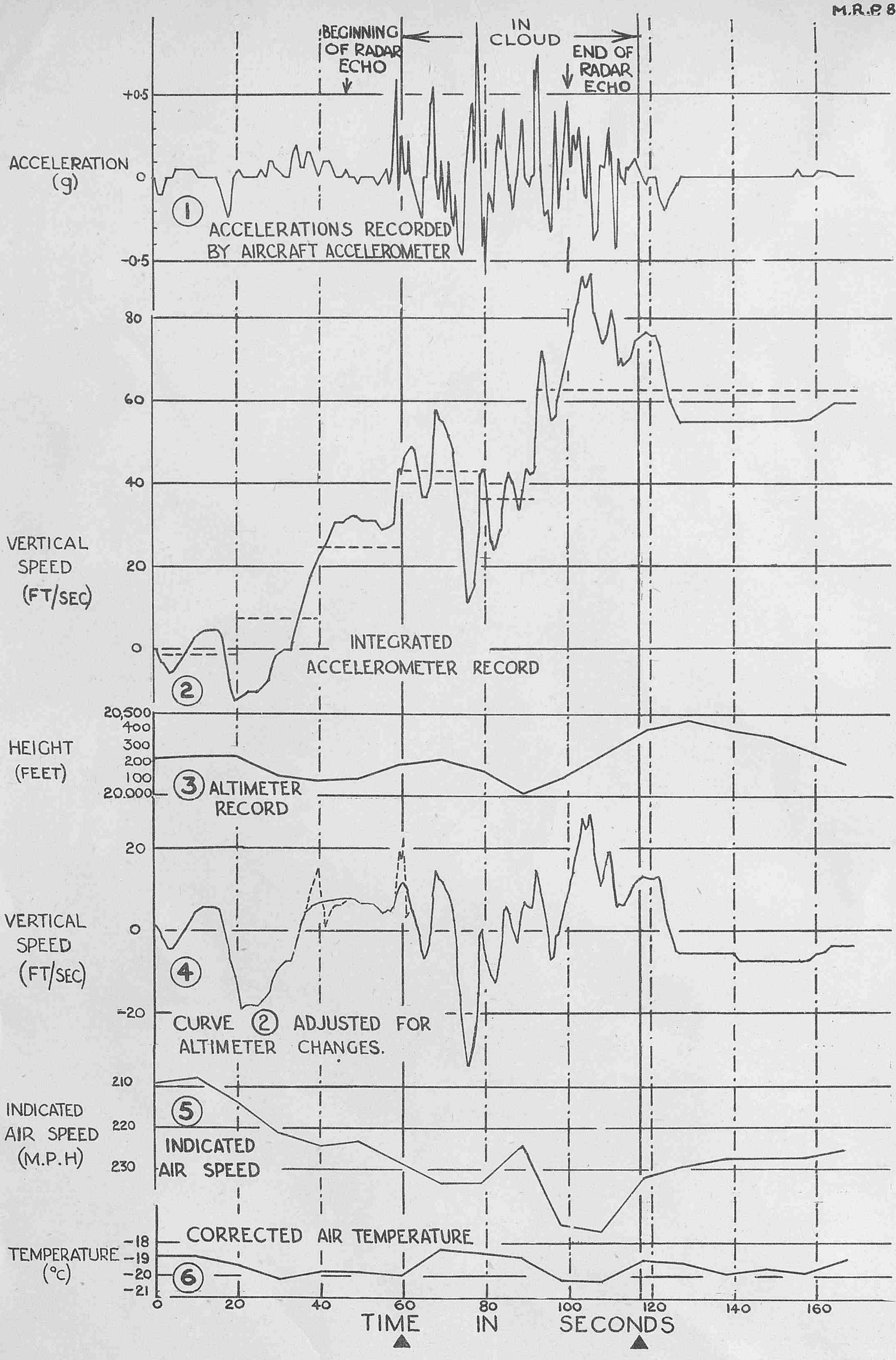


FIG. III. CUMULONIMBUS INVESTIGATION - 13TH JUNE 1952 - 20,200'.

M.R.P. 820.

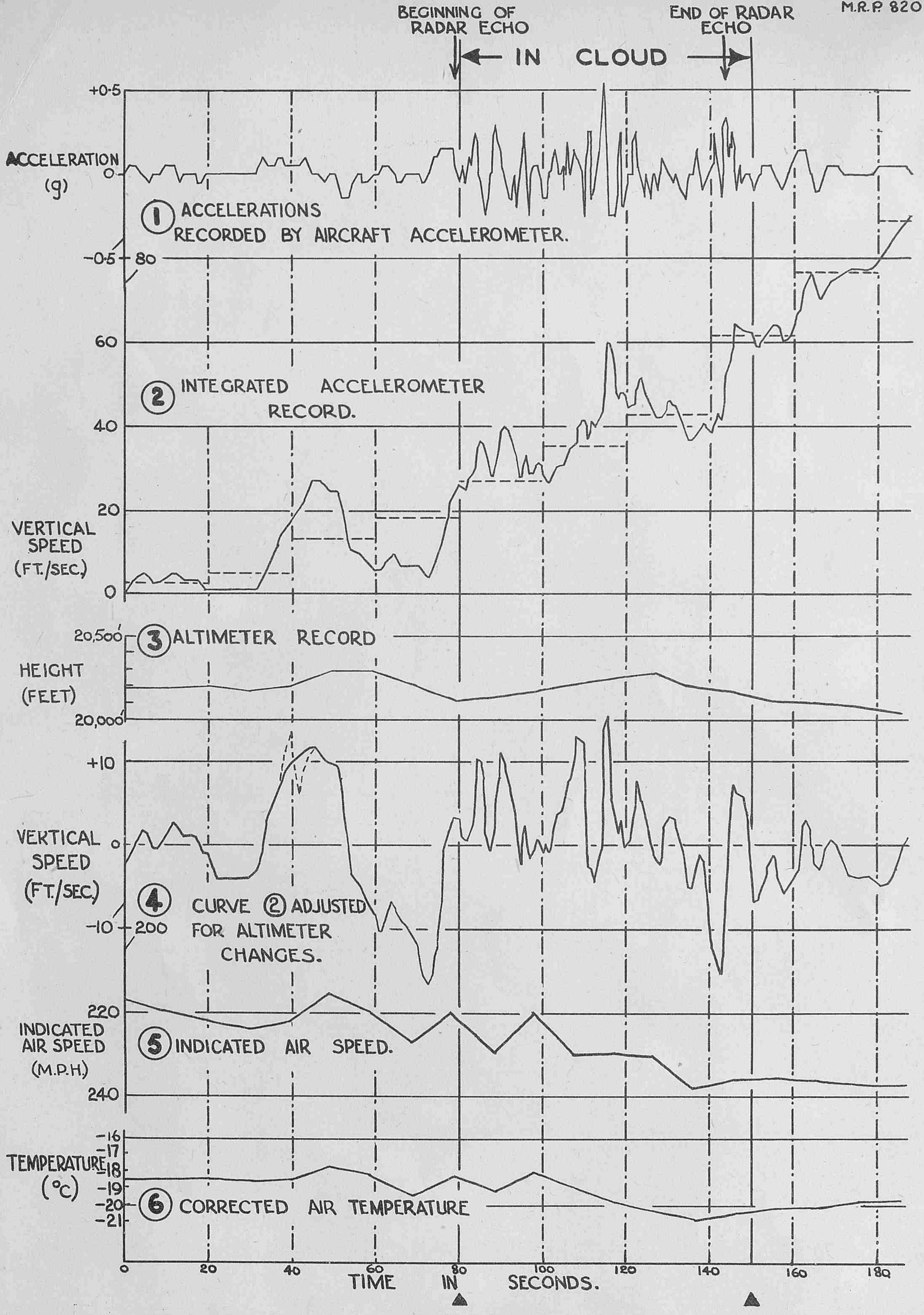


FIG. IV. CUMULONIMBUS INVESTIGATION-13TH JUNE 1952-RUN 4- 20,200'

M.R.P. 820.

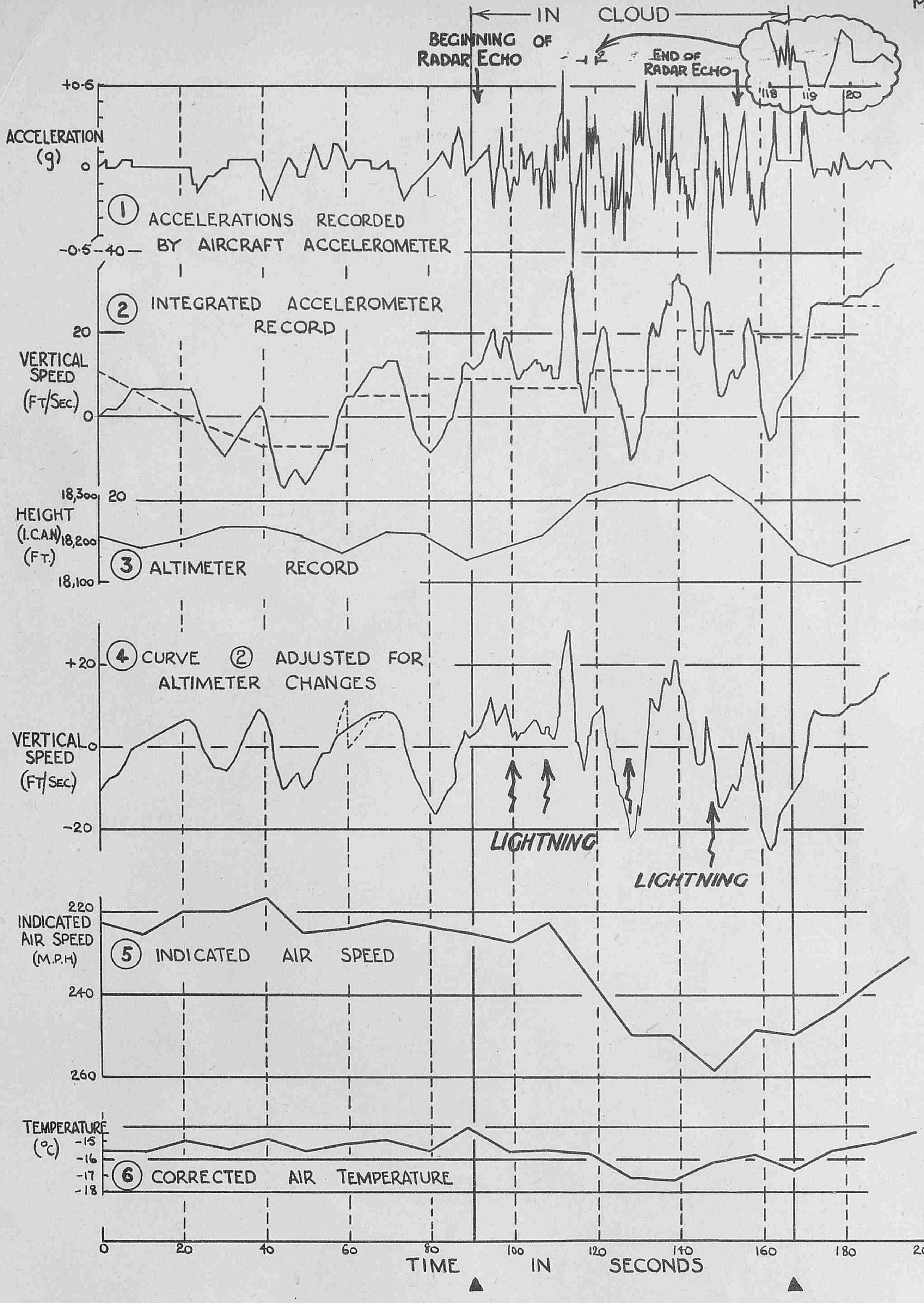


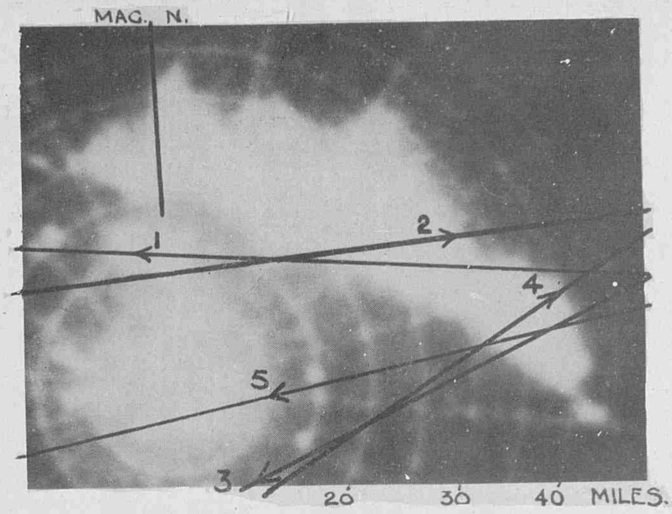
FIG. V. CUMULONIMBUS INVESTIGATION - 13TH JUNE 1952 - RUN 5 - 18,200'.

M.R.P. 820.

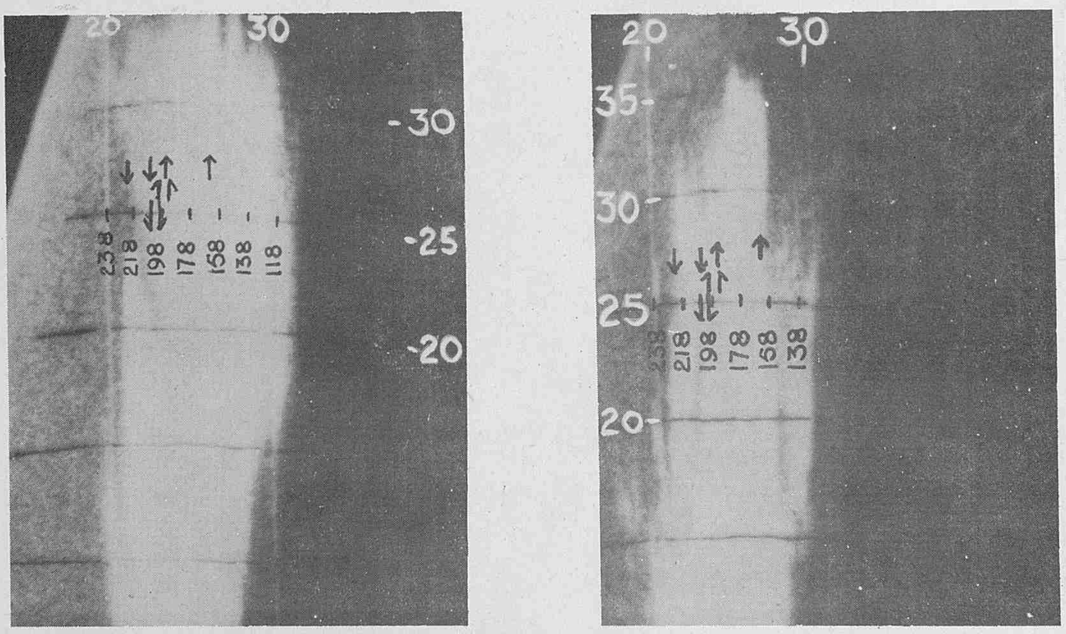
# RADAR ECHO PHOTOGRAPHS FOR THUNDERSTORMS OF 13TH JUNE 1952

IN HEIGHT/RANGE PHOTOGRAPHS ↑, ↓ INDICATE UP AND DOWN GUSTS & ⇈, ⇊ UP AND DOWN CURRENTS.

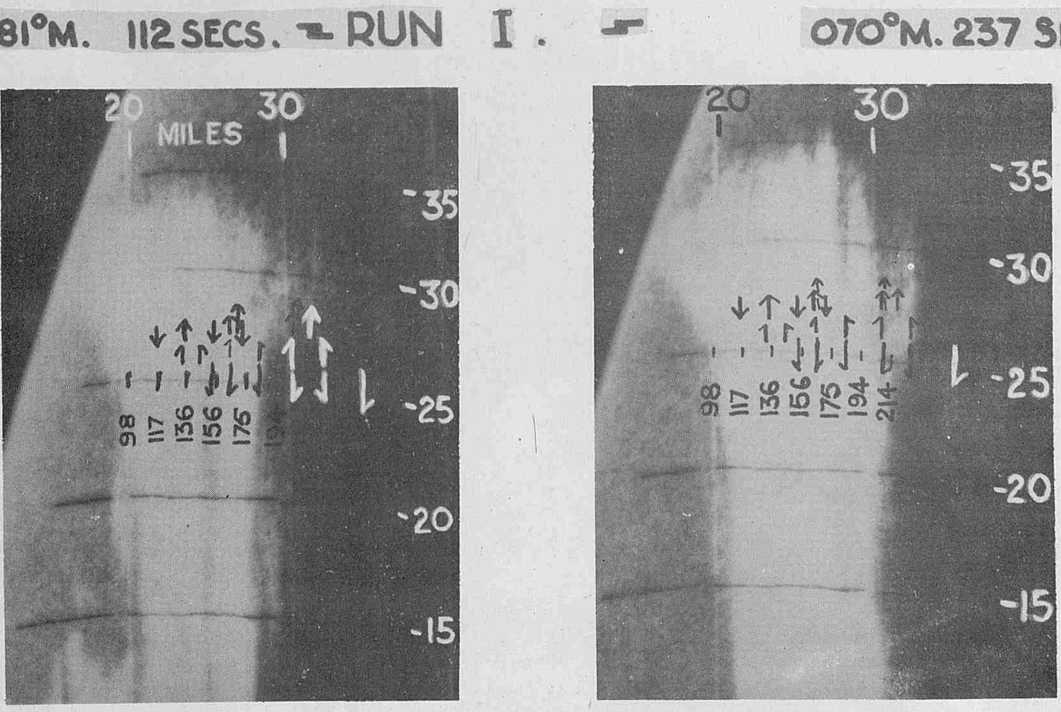
ON PHOTOGRAPHS RANGES ARE GIVEN IN STATUTE MILES AND HEIGHTS IN THOUSANDS OF FEET.


MAG. N.
1
2
3
4
5
20
30
40
MILES.

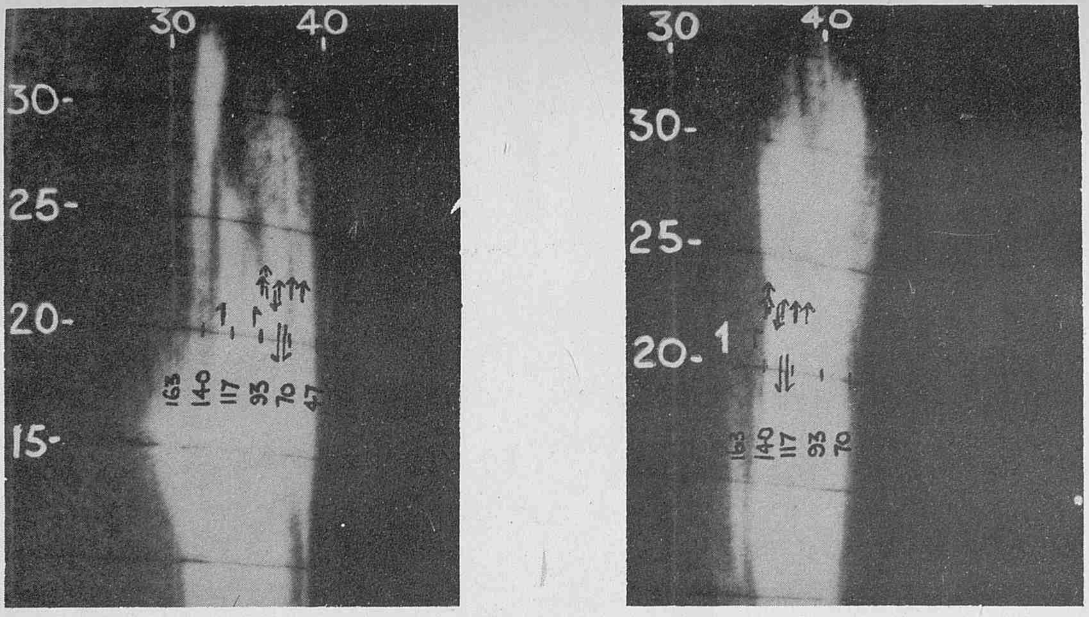
P.P.I. PHOTOGRAPH TAKEN 3 MINUTES AFTER RUN 5. THE TRACKS OF THE AIRCRAFT ARE SHOWN.


20
30
-30
-25
-20
238
218
198
178
158
138
118
20
30
35
30
25
20
238
218
198
178
158
138

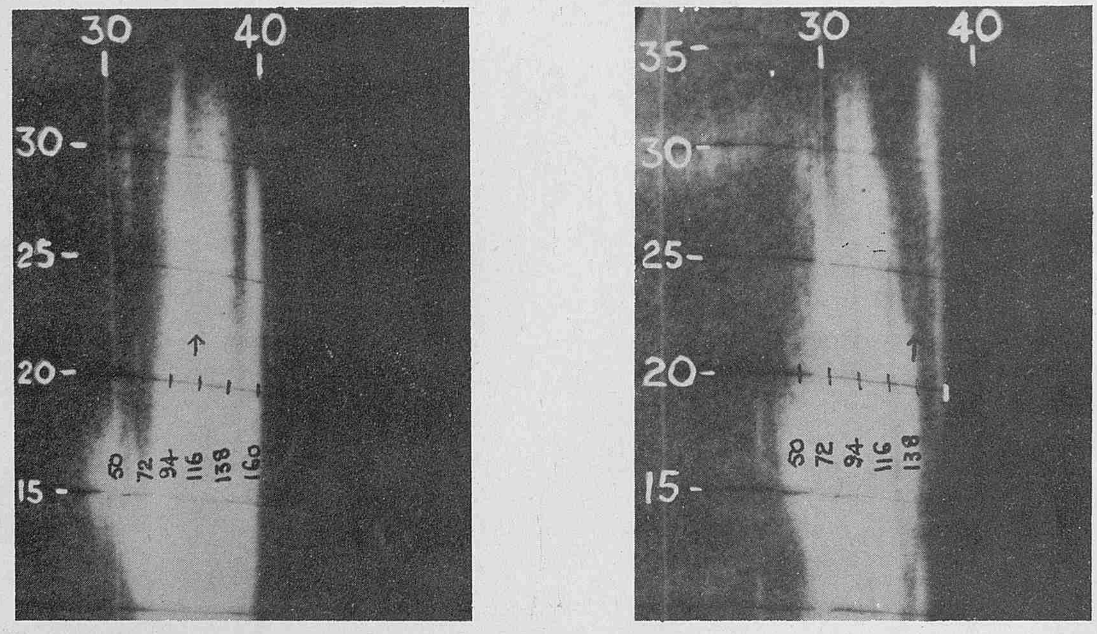
081°M. 112 SECS. – RUN I. – 070°M. 237 SECS.


20
MILES
30
-35
-30
-25
-20
-15
98
117
136
156
175
194
20
30
-35
-30
-25
-20
-15
98
117
136
156
175
194
214

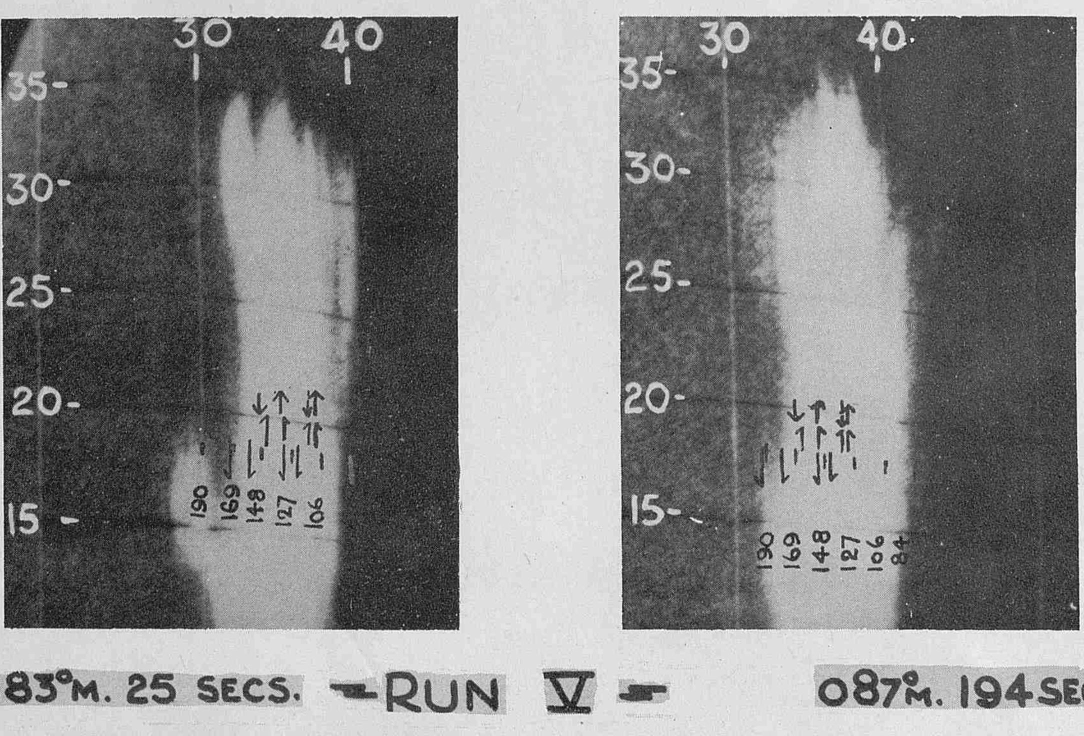
060°M. 129 SECS. – RUN II – 070°M. 185 SECS.



087°M. 50 SECS. — RUN III — 093°M. 163 SECS.



085°M. 69 SECS. — RUN IV — 081° 185 SECS.



083°M. 25 SECS. — RUN V — 087°M. 194 SECS.

MRP's 821, 822, 823, 826, 827

Control Number
C.J. 8053

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M.R.P. 821

S.C.II/155

10th June, 1953.

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## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# Note on the structure of the high altitude strong wind belt in the Middle East in winter

by

J. K. Bannon (Meteorological Office)

## Summary

From an analysis of the upper air observations at Habbaniya and Bahrein for January 1951 a meridional cross-section is constructed showing the mean structure for that month of the strong wind belt in the upper troposphere over Iraq. The analysis is subjective and liable to errors, but the self-consistency of the results indicates some reliability.

## 1. Introduction

The strong winds which blow in the upper troposphere over the Middle East in winter and spring are well known. Frost(1) and Gilchrist(2) have shown meridional cross-sections of the mean flow of the air over Iraq in January and April. This strong wind belt can certainly be associated with the polar front. It is also probably in part a continuation of a strong flow which is found in the upper troposphere in many longitudes between Lats. 25 and 40°N (2 and 3). The structure of this strong wind belt, whether it resembles a jet-stream of more temperate latitudes or is less sharply defined, is in some doubt.

The following note describes an attempt to discover the structure of this stream from a study of the observations at Habbaniya (33°22'N 43°34'E) and Bahrein (26°14'N 50°25'E) twice daily in January 1951.

## 2. Method of analysis

Temperature and wind data for Habbaniya and Bahrein for the levels 400 mb. and upwards were plotted for each hour of observation in January 1951 as shown in the examples in Fig. 1 which give data for 12 - 14th January. Mean wind direction between 300 and 100 mb. is given below each mean plot; variation about this mean was usually less than 15 to 20° - wind direction seldom changes much with height at those levels in the Middle East. Tropopause levels were determined using the experimental definitions given in reference (4).

From the observations plotted as in Fig. 1 an estimate was made of the plan position of the axis of strongest winds relative to the two observing stations. Fig. 2 shows the method of classifying the position of the axis. The positions are recorded on Fig. 1 for the relevant observations. In doing this, consideration was given to continuity and to the rise or fall of

/temperature

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temperature in the middle troposphere as indicating northward or southward motion respectively. The estimates are, therefore, subjective and must be liable to considerable error on occasions. It should be mentioned that the analyst may well have been influenced by a knowledge of the typical variations in jet-streams of temperate latitudes, but since some of the results of the analysis are dissimilar to those for a jet-stream this source of error may not be serious.

From each pair of observations an estimate was made of the minimum width of that part of the stream with a speed of 100 kt. or more. For example (see Fig. 1) the minimum width of the strong winds at 14 G.M.T. 14th was estimated as 400 mi., allowing for wind direction and the fact that the wind was appreciably greater than 100 kt. at both Bahrein and Habbaniya. Such estimates were largest when the axis of the stream was north of Habbaniya (zone (VI)), the wind then often exceeding 100 kt. at both stations. This result is to be expected if the axis of the strong winds is nearer the polar edge of the stream. The mean value of such minimum widths was 520 mi. for the 21 cases in zone (VI).

Next, the observations at the two stations were collected together according to the position of the axis and means were formed of the parameters plotted in Fig. 1. In the first instance this was done for wind observations only, but, as the results appeared encouraging, temperatures and tropopause levels were also averaged. On many occasions a level of maximum wind was not observed as wind was still increasing with height when observations ceased as the radar target passed out of range. To take some account of these, a minimum mean height and speed of the maximum wind was computed for each zone by taking the last wind observation as the maximum wind when wind was still increasing with height.

All these mean values were then plotted as shown in Fig. 3. If we assume that northward or southward fluctuations of the position of the stream do not affect its mean characteristics, (an assumption which is probably not fully justified; see below) then we may consider the mean values plotted in Fig. 3 as referring to stations at fixed intervals across a mean stream. Habbaniya is approximately 400 mi. north of Bahrein and the mean wind in the upper troposphere over Iraq was approximately westerly in January 1951, so that the interval between the plots may be taken as about 100 mi.

Isopleths of mean wind speed have been drawn on Fig. 3. Loss of observations at the higher levels makes some of the plotted mean values at 200 mb. and above unrepresentative and some allowance has been made for this in drawing the isokinetics. The minimum heights and speeds of the mean maximum wind computed as described above were also used in assessing the position of the central isopleths.

Isopleths of potential temperature were also drawn on Fig. 3 from the mean temperatures plotted. Like the isokinetics they appear to follow a pattern and the slopes of these lines were used to reassess levels of maximum wind assuming the geostrophic relation and hence to make minor corrections to the isopleths of mean wind.

## 3. Discussion

It is perhaps surprising that the subjective analysis described in Section 2 should produce such a self-consistent picture as Fig. 3. This consistency probably indicates that on the average the assessments of position of the axis of the stream were not far out and that the assumptions made in the analysis were mainly valid. It will be noted, however, that there is only limited agreement between the observations at Habbaniya in zones (V) and (VI) and Bahrein in zones (I) and (II) where the two sets should overlap in Fig. 3. This is some indication that the stream differs in detail (temperatures, tropopause levels, etc.) according as it is north or south of its mean position, though it does not necessarily show that the structure of the stream varies with position.

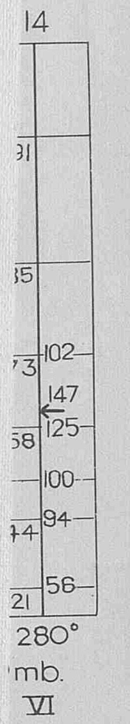
Another discrepancy in the analysis is that though each pair of observations was classed as in Fig. 2 the axis of the stream in Fig. 3 appears not at Habbaniya (V) but nearer Habbaniya (VI) and Bahrein (II). This fault is not considered serious, in view of the roughness of the classification.

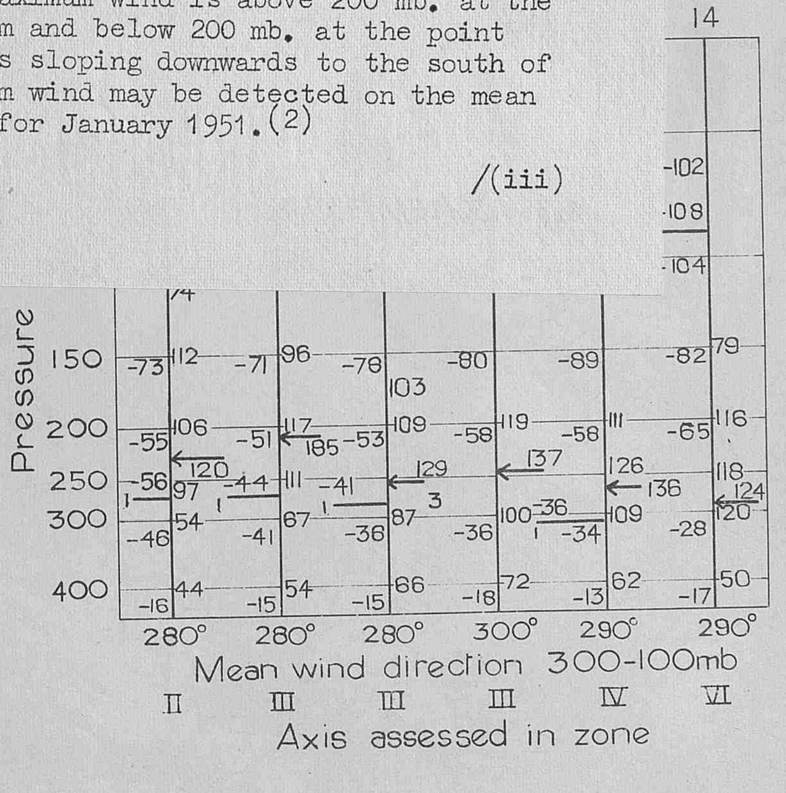
Points of interest are

(i) The stream appears to be wide and without the strong horizontal gradients of wind speed which are found in typical jet-streams of higher latitudes. The 100 kt. isokinetic is seen to be about 700 mi. wide, i.e. the mean width of the strong stream (100 kt. or more) is almost 10° of latitude. That this estimate is not greatly in excess of the true value is shown by the mean lower bound of 520 mi. for occasions in zone (VI) (see Section 2). There remains the doubt, however, whether this method of analysis is adequate to show up a really strong horizontal shear occurring on the polar side of the stream.

(ii) The level of maximum wind falls southwards through the strong wind belt. This is the reverse of what is found in the typical frontal jet-stream of middle latitudes. It may be argued that this result may be incorrect, because more ascents failed to reach the level of maximum wind in the stronger winds on the right-hand side of Fig. 3. This effect must certainly be present, but some allowance has been made for this in drawing the isopleths and the weight of the evidence is for the result as stated. For example, there is little doubt that the level of maximum wind is above 200 mb. at the centre of the stream and below 200 mb. at the point Bahrein (IV). This sloping downwards to the south of the level of maximum wind may be detected on the mean meridional section for January 1951.(2)

/(iii)





(iii) The upper tropopause does not appear to vary greatly; it is consistently higher at Bahrein than at Habbaniya.

(iv) The lower tropopause appears to slope upwards towards the axis of strong winds but the evidence for this on the south side of the axis is slender.

## 4. Conclusions

Many criticisms may be made of the above rough analysis. If the results in Fig. 3 are accepted because of their self-consistency, it appears that in January 1951 the strong wind belt over Iraq and the Persian Gulf was in the mean a broad stream without strong horizontal gradients; its axis was about 190 mb. and the breadth of the highest speeds (100 kt. or more) was about 700 miles.

## References

| | | |
|---|---|---|
| (1) | Frost, R. | The upper air circulation in low latitudes and its relation to certain climatological discontinuities. Met. Res. Pap. 706, 1952. |
| (2) | Gilchrist, A. | Upper winds in the Tropics and Sub-Tropics. Met. Res. Pap. 795, 1953. |
| (3) | Namias, J. and Clapp, P.F. | Confluence theory of high tropospheric jet-stream. J. of Met., 6, 1949, p. 330. |
| (4) | - | Upper air data, 1946-50, Habbaniya. M.O. 555, Part 3, 1953. |

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position of the central isopleths.

Isopleths of potential temperature were also drawn on Fig. 3 from the mean temperatures plotted. Like the isokinetics they appear to follow a pattern and the slopes of these lines were used to reassess levels of maximum wind assuming the geostrophic relation and hence to make minor corrections to the isopleths of mean wind.

## 3. Discussion

It is perhaps surprising that the subjective analysis described in Section 2 should produce such a self-consistent picture as Fig. 3. This consistency probably indicates that on the average the assessments of position of the axis of the stream were not far out and that the assumptions made in the analysis were mainly valid. It will be noted, however, that there is only limited agreement between the observations at Habbaniya in zones (V) and (VI) and Bahrein in zones (I) and (II) where the two sets should overlap in Fig. 3. This is some indication that the stream differs

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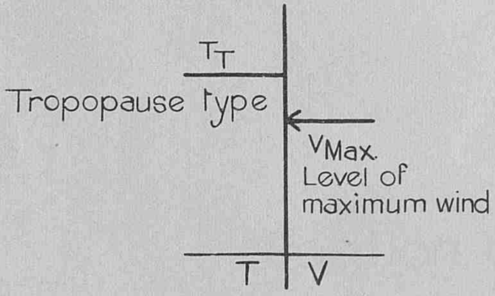
# Fig. 1

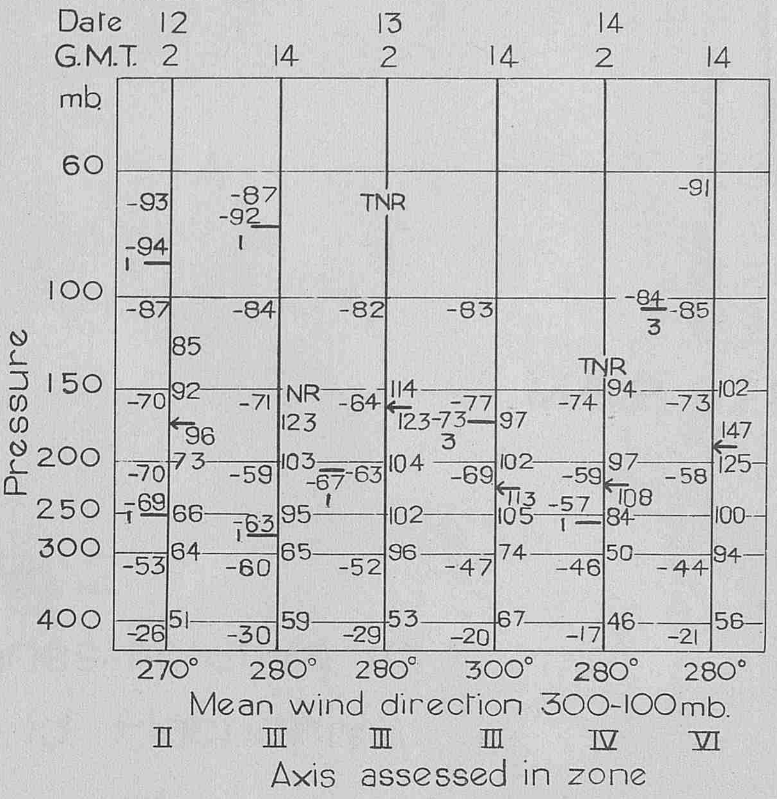
## Upper winds and temperature at Habbaniya and Bahrein 12-14th January, 1951

PLOTTING KEY

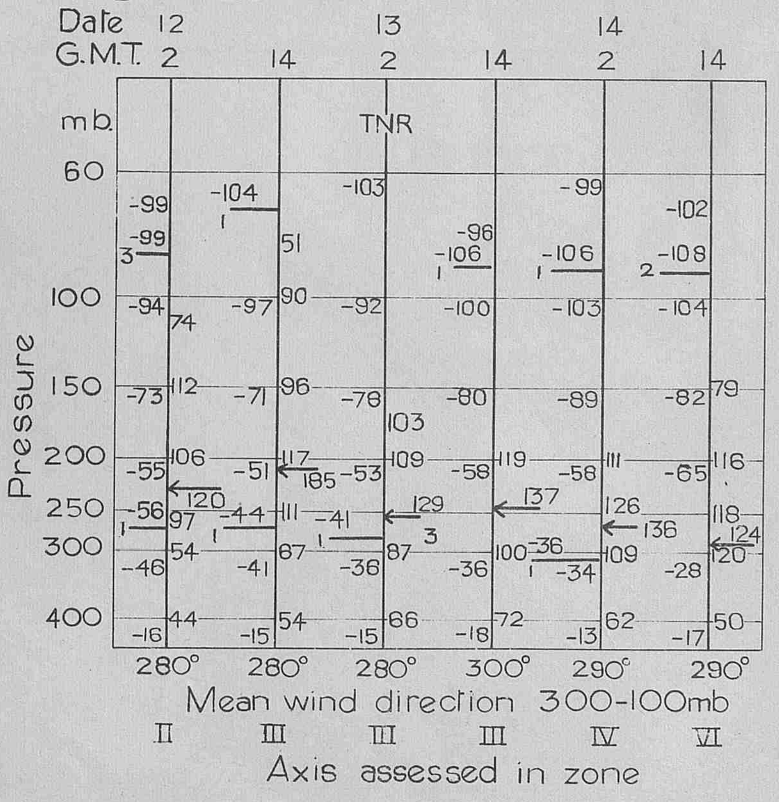


$T_T$
Tropopause type
$V_{Max.}$
Level of maximum wind
T
V

HABBANIYA



BAHREIN



T = Temperature °F.

$T_T$ = Temperature at the tropopause

V = wind speed (kt.)

$V_{Max.}$ = Maximum wind speed

TNR = Upper tropopause not reached

NR = Level of maximum wind not observed

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Fig. 2
Position of zones relative to Bahrein and Habbaniya

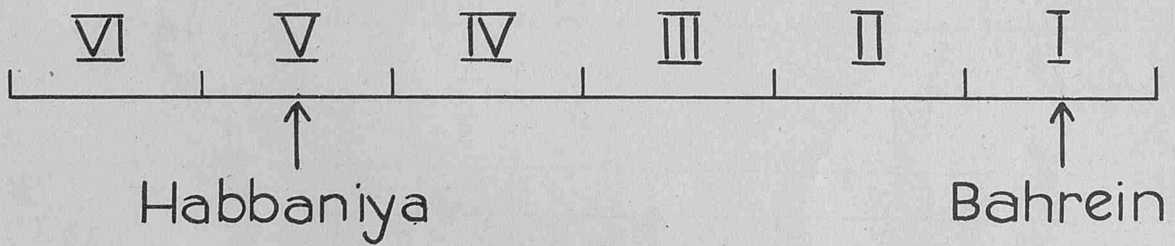
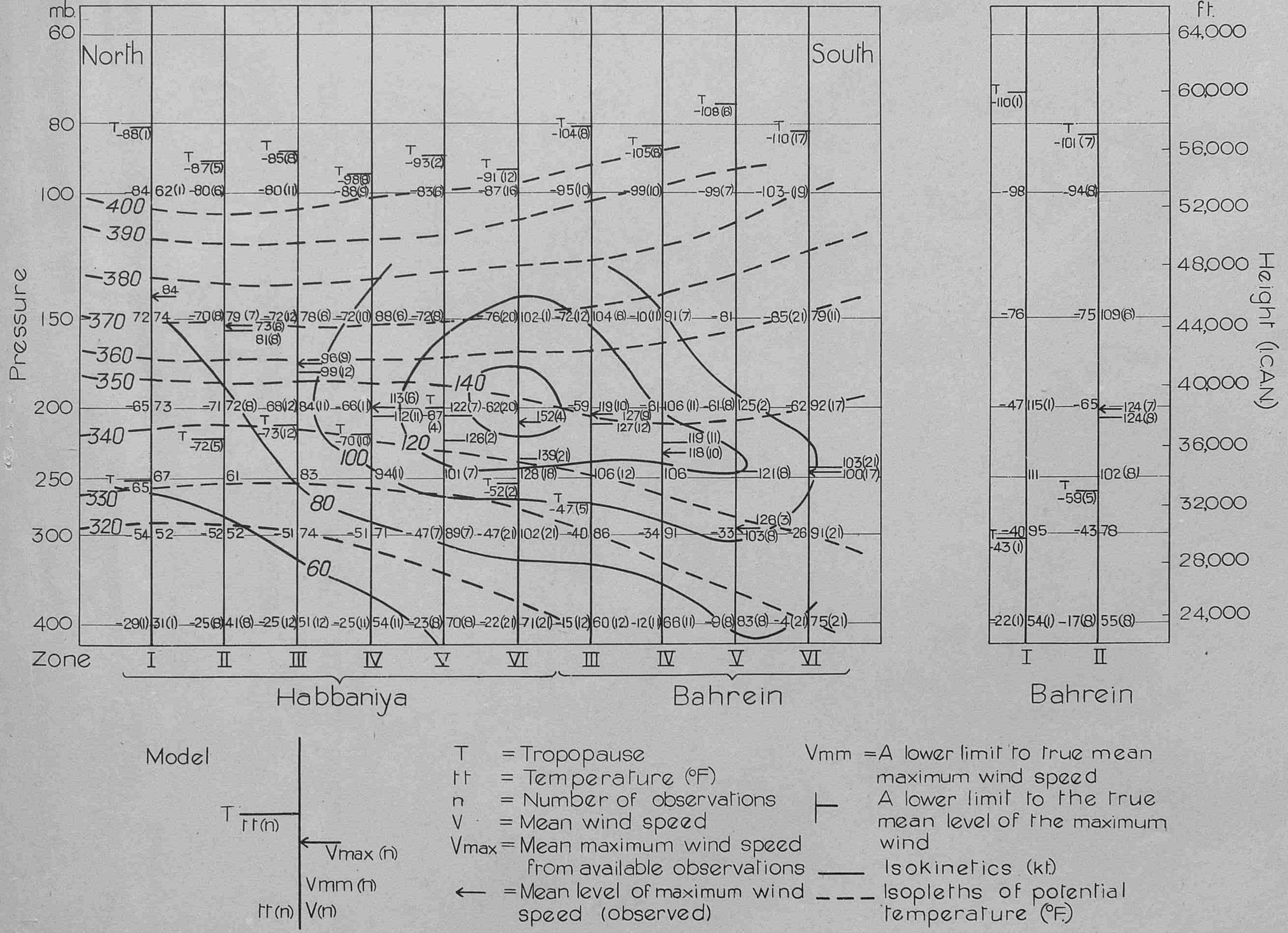

VI
V
IV
III
II
I
Habbaniya
Bahrein

Fig. 3
Constructed mean cross-section for 45°E
January, 1951



At the time of issue this report has not
been published. Any author desiring to
quote this paper should refer to it as "A
paper of the Meteorological Research
Committee, London,
M.R.P. 822, 1953 by Mr. R.J. Murgatroyd;
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M.R.P. 822.

S.C.I./79.

16th June, 1953.

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## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# A Report on the Determinations of the Speed Correction Coefficients of Aircraft Thermometers made by M.R.F. in 1951

R.J. Murgatroyd (Meteorological Office)

## Summary

In Part I recent flight determinations of the speed correction coefficients of three flat plate and two conical head aircraft thermometers are described. The experiments were made at 5000 and 21000 feet with indicated airspeeds of 120 and 180 knots. No consistent variation of these coefficients with altitude was found.

The technique of these experiments is discussed in Part II. It is considered that the scatter of individual results in this work will invalidate attempts with the present aircraft available to improve substantially the accuracy of temperature measurement in aircraft.

## Part I. Recent Measurements of Speed Correction Coefficients

### Introduction

Flight determinations by the Meteorological Research Flight of the speed correction coefficients of the M.O. flat plate resistance thermometer and the R.A.E. pitot head resistance thermometer were reported in M.R.P. No. 527 (1) where it was stated that in both cases these coefficients decrease with increasing height.

The possible causes for this were examined in M.R.P. No. 677 (1) and it was concluded then that in the case of the flat plate thermometer this variation was probably due to a change of the position of transition from laminar to turbulent flow on the thermometer surface. A larger proportion of laminar flow would lead to a lower speed correction coefficient so that if the flow is more laminar at higher altitudes the above variation would be explained.

An attempt was made to verify this idea by flight determinations in a Hastings aircraft using the following thermometers:-

1. A standard M.O. flat plate thermometer.

2. A flat plate thermometer with two trip wires arranged ahead of the element to ensure that it was always in their turbulent wake.

3. Another flat thermometer with [illegible]

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 822 by Mr. R.J. Murgatroyd A copy is available in the Library of the Meteorological Office."

M.R.P. 822.
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An attempt was made to verify this idea by flight determinations in a Hastings aircraft using the following thermometers:-

1. A standard M.O. flat plate thermometer.
2. A flat plate thermometer with two trip wires arranged ahead of the element to ensure that it was always in their turbulent wake.
3. Another flat thermometer with suction slots connected to a Venturi tube arranged as far as possible to ensure that the element was always in a laminar flow.

/4.

4. A Sangamo-Weston conical head thermometer.

5. A Meteorological Office experimental conical head thermometer.

The use of conical head thermometers was also suggested in M.R.P. No. 677.

## Flight procedure

The thermometers were mounted at various points under the nose of the Hastings aircraft and connected to M.O. thermometer bridges. Three or four thermometers were read simultaneously every ten seconds on each run together with height and airspeed. Twenty such readings were taken at the 180 knots airspeed, and thirty at the 120 knots airspeed to ensure an equal distance of air traversed on each run. Runs were made alternatively in reciprocal directions between two air positions in the order 120 knot run, 180 knot reciprocal run, 180 knot run, 120 knot reciprocal run. The averages of the 180 and 120 knot legs were then used in the subsequent calculations. Usually three or four of these series would be flown in succession, the overlapping 180, 120, 120, 180 knot runs also being used as a sequence to calculate a speed correction coefficient. Determinations were made at two levels, 5000 feet and 21000 feet. In the calculations corrections were applied for calibration compressibility errors but not position errors. The reasons for this are given in Part II of this report.

The flights were divided into two series, the M.O. flat plate thermometer being used in both series. The trip wire was added to the second flat plate thermometer between Series 1 and Series 2.

The results of the calculations are given in Table II.

## Summary of results

The principal results are as follows:-

1. In the first series the M.O. flat plate speed correction coefficient decreased from 1.73 to 1.71 between 5000 and 21000 feet.

   The corresponding decrease in the second series was 1.66 to 1.62.

   It is not known why there was a decrease from 1.73 to 1.66 between the two series as the apparatus was not changed in any way.

   The speed coefficient of the second flat plate in Series 1 increased from 1.50 to 1.60 over this height range.

2. The speed coefficient of the flat plate with trip wires decreased from 1.16 to 1.14 and that of the slotted flat plate increased from 1.59 to 1.60.

3. The speed coefficient of the two conical head thermometers increased with height, the Sangamo-Weston instrument from 1.69 to 1.74 and the M.O. instrument from 1.83 to 1.87.

4. The standard deviation of individual measurements was high, the greatest being 0.16 and the smallest 0.04. The range between smallest and greatest single measurements varied between about 0.6 and 0.2. In each case the scatter at 5000 feet was considerably greater than that at 21000 feet.

/Discussion

## Discussion

The above results do not substantiate the previous finding that the speed correction coefficient of the M.O. flat plate thermometer decreases with height. It may be that this in fact does happen at greater heights and speeds than can be obtained by the Hastings aircraft but within these present limitations it is evident that this variation is of no practical importance as it is always masked by the scatter of the individual determinations.

The conical head thermometers showed a slight increase of the speed correction coefficient with height, a result which was anticipated in M.R.P. No. 677.

The experiment failed to demonstrate whether the trip wires or slotted elements eliminated any effects due to the position of changeover from laminar to turbulent flow on the thermometer surface as the M.O. flat plate thermometer with which they were compared did not show a significant change of its coefficient with height. It did demonstrate however that the coefficients vary considerably with the position of the element on the aircraft, and it seems that for work of good accuracy any installation would need a separate calibration before use.

The main result of the trip wires is to reduce the coefficient considerably. Cooling effects due to a bending of the streamlines of air flow have been noted in the literature and rather analogous effects are used in Vortex thermometers. It may be that development along the lines of reducing the coefficient itself may be more profitable for general service use than directing the effort into improving its constancy.

The accuracy of this work and its dependence on the accuracy of the other aircraft instruments is discussed in Part II of this report.

Table I. Flight results

1st Series

| Type of Thermometer | Height in feet | No. of measurements | Mean α | Standard deviation | Smallest α measured | Greatest α measured |
|---|---|---|---|---|---|---|
| Standard M.O. flat plate | 5000 | 20 | 1.73 | 0.12 | 1.53 | 1.96 |
| | 21000 | 16 | 1.71 | 0.07 | 1.61 | 1.86 |
| Sangamo-Weston conical head | 5000 | 22 | 1.69 | 0.16 | 1.36 | 1.92 |
| | 21000 | 18 | 1.74 | 0.11 | 1.48 | 1.90 |
| Flat plate without trip wires | 5000 | 22 | 1.50 | 0.10 | 1.26 | 1.62 |
| | 21000 | 18 | 1.60 | 0.04 | 1.53 | 1.68 |

2nd Series

| Type of Thermometer | Height in feet | No. of measurements | Mean α | Standard deviation | Smallest α measured | Greatest α measured |
|---|---|---|---|---|---|---|
| Standard M.O. flat plate | 5500 | 14 | 1.66 | 0.07 | 1.55 | 1.78 |
| | 21000 | 15 | 1.62 | 0.06 | 1.43 | 1.69 |
| M.O. conical head | 5500 | 14 | 1.83 | 0.08 | 1.67 | 1.96 |
| | 21000 | 15 | 1.87 | 0.07 | 1.81 | 2.04 |
| Flat plate with trip wires | 5500 | 14 | 1.16 | 0.07 | 1.04 | 1.24 |
| | 21000 | 15 | 1.14 | 0.03 | 1.06 | 1.17 |
| Slotted flat plate with Venturi | 5500 | 14 | 1.59 | 0.10 | 1.36 | 1.72 |
| | 21000 | 15 | 1.60 | 0.03 | 1.55 | 1.64 |

/Part II.

## Part II. The Technique of Determining the Speed Correction Coefficients of Aircraft Thermometers

### 1. Definition of speed correction coefficient and its standard method of determination

On p.15 of M.O. 470 (5) the correction $\Delta T$ to be subtracted from the indicated dry-bulb temperature $T_i$ is given by

$$\Delta T = \alpha \left( \frac{V}{100} \right)^2 \qquad (1)$$

where $V$ is the true air speed and $\alpha$ is the speed correction coefficient. The recommended method of determining $\alpha$ is to fly in suitable conditions at a constant pressure level taking readings of indicated temperatures at speeds $V_1$, $V_2$, $V_3$, $V_4$, $V_3$, $V_2$, $V_1$ in order, where $V_1$, $V_4$ are minimum and maximum aircraft speeds respectively. $\alpha$ is then given by the slope of the best straight line drawn through values of indicated temperature $T_i$, plotted against $(V/100)^2$.

The value of true airspeed V is in each case obtained from the equivalent airspeed $V_i$ by the equation:-

$$\left(\frac{V}{V_i}\right)^2 = 1.95 \frac{T + 459}{P} \qquad (2)$$

where T, P, are the ambient temperature in degrees Fahrenheit and pressure in millibars respectively and $V_i$ is the reading of the air speed indicator with various corrections applied. These corrections increase in importance with speed and altitude and are discussed below.

It is also necessary to apply corrections to the altimeter readings before they can be used in equation (2) and these corrections are also discussed in a subsequent paragraph.

### 2. The accuracy required in the value of the speed correction coefficient

In the work described in Part I of this report $\alpha$ was found to be about 1.7 when the temperature was measured in degrees Fahrenheit and the airspeed in knots. The value of $\Delta T$ °F and the error in $\Delta T$ if $\alpha$ is 0.1 in error are given for various airspeeds in Table II.

Table II

Some values of speed corrections and their errors at various airspeeds.

| Airspeed<br>V Knots | Correction<br>$\Delta T$ °F | Error in<br>$\Delta T$ °F |
|---|---|---|
| 100 | 1.7 | 0.1 |
| 200 | 6.8 | 0.4 |
| 300 | 15.3 | 0.9 |
| 400 | 27.2 | 1.6 |
| 500 | 42.5 | 2.5 |
| 600 | 61.2 | 3.6 |

It appears from this Table and the results of Part I that errors of 3.5°F or more may arise in temperature readings from fast aircraft due to uncertainties in the value of the speed correction coefficient. In the relatively slow Hastings aircraft used by M.R.F. the error is normally between 0.1 and 0.3°F.

## 3. Computation technique when the speed range is small

In para. 1 the technique involves using a range of airspeeds say $V_1$ to $V_4$. As the available indicated speed range was only 120 to 180 knots it was considered better to concentrate the readings at the corrected lowest, $V_L$, and the highest, $V_H$, equivalent airspeeds with respective indicated temperatures $T_L$, $T_H$. It follows then from equations (1) and (2) that providing both high and low speed runs are made at the same ambient pressure P then:-

$$\alpha = \frac{(T_H - T_L)\,P}{1.95\left\{\left(\frac{V_H}{100}\right)^2\left(T_L + 459\right) - \left(\frac{V_L}{100}\right)^2\left(T_H + 459\right)\right\}} \qquad (3)$$

$$(T + 459) = \frac{\left(\frac{V_H}{100}\right)^2\left(T_L + 459\right) - \left(\frac{V_L}{100}\right)^2\left(T_H + 459\right)}{\left(\frac{V_H}{100}\right)^2 - \left(\frac{V_L}{100}\right)^2} \qquad (4).$$

For a small speed range this technique has statistical advantages and also simplifies the computation. It does however assume that equation (1) is correct. For large speed ranges it may be possible to detect deviations from this equation and then a range of speeds would have to be traversed.

## 4. The measurement of temperature

Care must be taken to provide the thermometer with a good exposure in a position under the wing or under the nose of the aircraft. All flying should be carried out without the sun's radiation falling directly on the element. The thermometer is calibrated in still air in the laboratory before use.

The problem is then to find a representative mean temperature for the same sample of air at the different aircraft speeds. This is complicated by (a) small scale inhomogeneities in the air mass (b) variation of mean temperature with time and (c) variation of mean temperature with distance. These are now considered separately:-

(a) Very small scale inhomogeneities of the type described in M.R.P. No. 658 (1) will be averaged by the lag effects of most types of aircraft thermometers. Those of the types described in M.R.P. No. 376 (1) however can only be minimised by flying for at least 3 to 5 minutes on each run i.e. up to 20 air miles and taking readings at least every 10 sec. In addition each run should be made as accurately as possible between two fixed air positions. High speed runs will then be of shorter duration than low speed runs.

(b) The change of mean temperature with distance is to a large extent covered by flying on reciprocal bearings between the two air positions. It may be some advantage to arrange the aircraft's course to be approximately along the expected direction of the isotherms at the level concerned but in practice this often conflicts with other considerations and is rarely carried out.

(c) Advective change of temperature is also covered by using air positions as above. Other possible changes of mean temperature with time will be minimised by keeping the period of each determination small. To examine the importance of this effect a small subsidiary investigation was carried out on the changes of temperature measured by radiosonde at Larkhill between 0900 and 1500 hours during the period Jan. 1950 to Sept. 1951 with the following result.

Table III

Mean temperature changes (neglecting sign) 0900-1500 hours at 850 and 450 mb at Larkhill.

| Level mb | Average temp. change °F | Maximum temp. change °F |
|---|---|---|
| 850 | 2.5 | 16 |
| 450 | 3.1 | 19 |

It thus appears that at least 0.5°F changes are to be expected in an hour and up to 3°F may easily be possible. A large proportion of these are advective but it seems that a change of 0.5 to 1°F due to dynamic and radiation effects is quite possible.

The measured changes due to the speed range are about 3°F at 5000 feet and 6°F at 21000 feet. Hence it is inadvisable to spend more than 15 to 30 minutes on a single experiment.

Another consideration which is discussed in greater detail below is the effect of varying altitude on the temperature readings. A lapse rate of 0.5°F in 100 feet is common and increases of greater magnitude are found in inversions.

## 5. The measurement of height and airspeed

(i) The altimeter and airspeed indicator are standard aircraft instruments operated by the air pressure in the pitot and static sources. The instruments themselves are calibrated in the laboratory against known pressures to determine their instrumental errors but a knowledge of the behaviour of the pitot and static sources requires a calibration of the aircraft itself.

The principal corrections which then arise in use are:-

(a) Calibration compressibility correction to the airspeed indicator readings.

(b) Corrections for position errors. Of these the static position error which affects both the height and airspeed readings is the more important.

At high speeds the static position error is varied by compressibility effects.

(ii) The calibration compressibility error arises from the fact that the airspeed indicator is calibrated as above for surface conditions only and at greater heights deviations from this calibration take place. A short Table is given below showing the magnitude of this error.

Table IV

Some calibration compressibility errors (knots)

| Indicated airspeed (knots) | Height in feet 5,000 | 10,000 | 20,000 | 30,000 | 40,000 |
|---|---|---|---|---|---|
| 120 | -0.1 | -0.2 | -0.5 | -1.1 | -2.1 |
| 180 | -0.5 | -0.7 | -1.8 | -3.7 | -6.7 |
| 300 | -1.7 | -3.3 | -8.2 | -15.3 | - |

Corrections were made for these errors in the work described in Part I of this report.

(iii) Static position errors are due to the static source of the aircraft not always registering correct static pressure and giving different values at different airspeeds and heights. Correction should be made to both airspeed and height readings for this error. Hence the recommended method of measuring speed correction coefficients in Ref. 3 is to predetermine the height correction to the altimeter reading at the various speeds at which the experiment is to be made and then arrange the flight plan so that all the speed runs are made at constant pressure height. Corrections are then also applied to the airspeed readings. Unless this is done the actual height of the aircraft will vary with the aircraft speed and the results of the experiment will be in error by an amount depending on the temperature lapse rate.

The Hastings aircraft used in this work was calibrated for position errors just above runway height by comparing the readings of the aircraft instruments with surface photographic determinations of the true height of the aircraft. The errors at other altitudes were then determined theoretically by the methods given in Ref. 2 in which the assumption is made that the static pressure varies with the Mach number according to a relation given by Glauert. This relation at present is considered the most satisfactory and can be used confidently within the operating conditions of the Hastings aircraft. Some values for the corrections determined for the position errors of this aircraft are given in Table V. Between 100 and 130 knots I.A.S. this aircraft is flown with 40° flaps and "clean" at greater speeds.

Table V

Some height corrections $\Delta h$ and speed corrections $\Delta V$ for position errors.

| Height feet | Speed knots 120 | | 180 | | 220 | |
|---|---|---|---|---|---|---|
| | $\Delta V$ | $\Delta h$ | $\Delta V$ | $\Delta h$ | $\Delta V$ | $\Delta h$ |
| | knots | feet | knots | feet | knots | feet |
| 5,000 | -1 | -14 | -1 | -42 | -3 | -95 |
| 10,000 | -1 | -17 | -2 | -49 | -4 | -110 |
| 20,000 | -2 | -23 | -3 | -69 | -6 | -156 |
| 30,000 | -3 | -35 | -4 | -103 | -8 | -234 |

(iv) It is possible however to ignore the effects of position errors and determine the speed correction coefficient accurately provided that the lapse rate throughout all the runs everywhere is approximately the wet-adiabatic. Under these conditions it so happens that the effect of position error on altitude opposes its effect on speed. It can be shown that when the fraction of full kinetic heating attained by the thermometer is the same fraction as the actual lapse rate is of the dry adiabatic lapse rate, then the two effects cancel. In practice both these fractions are about 0.7 when the lapse rate is wet-adiabatic. The work described in Part I and also that described in M.R.P. No. 527 was carried out as far as possible in these conditions and position error effects were then disregarded.

It should be emphasised however that under isothermal and inversion conditions large errors will be introduced by neglecting position errors in this way.

(v) In a large aircraft where there are several instrument positions a variation occurs between them and it is necessary always to use the same position in this work. These installation errors were examined in flight where four 5 minute runs were made at about 150 knots I.A.S. at 5000 feet and 21000 feet with three sets of instruments being read simultaneously every 10 seconds. The results are given in Table VI.

Table VI

Variation of height and airspeed readings in different aircraft positions.

| | 2nd Pilot's Instruments | | Navigator's Instruments | | Engineer's Instruments | |
|---|---|---|---|---|---|---|
| | I.A.S. knots | Height feet. | I.A.S. knots | Height feet. | I.A.S. knots | Height feet. |
| Run No. 1 | | | | | | |
| Mean | 151.1 | 5011 | 149.3 | 4997 | 149.9 | 4963 |
| Standard Deviation | 1.3 | 7.9 | 1.6 | 10.1 | 1.0 | 9.2 |
| Run No. 2 | | | | | | |
| Mean | 151.0 | 5010 | 150.0 | 4994 | 149.7 | 4966 |
| Standard Deviation | 1.0 | 9.5 | 1.1 | 10.5 | 0.9 | 11.9 |
| Run No. 3 | | | | | | |
| Mean | 151.5 | 20915 | 149.3 | 20957 | 149.9 | 20996 |
| Standard Deviation | 1.1 | 11.2 | 1.6 | 14.4 | 0.9 | 14.6 |
| Run No. 4 | | | | | | |
| Mean | 150.6 | 20908 | 149.3 | 20948 | 149.4 | 20997 |
| Standard Deviation | 1.9 | 14.1 | 2.2 | 20.4 | 2.1 | 11.4 |

It appears that variations of 20 to 40 feet in heights at 5000 feet and 80 to 100 feet at 21000 feet may easily be produced by reading different altimeters in the aircraft. Moreover different altimeters may read higher at different altitudes. The uncertainty in mean speed is between 1 and 2 knots.

/The

## The effect of systematic temperature and pressure differences introduced by flying the high and low speed runs at different levels

If the variation of temperature with altitude is $\partial T/\partial P$ °F per millibar and one run is flown at a pressure level $P_1$ and the other at a pressure level $P_2$, such that $P_1 - P_2 = \partial P$ the effective ambient temperature for one run will be T °F and for the other $T + \delta T$ °F.

To a good approximation equation (3) will then become:-

$$\alpha = \frac{T_H - T_L\left(1 + \frac{\delta T}{T}\right)}{\frac{1.95}{100^2}\left[\frac{V_H^2}{P_2}\left(T_L + 459\right)\left(1 + \frac{\delta T}{T}\right) - \frac{V_L^2}{P_1}\left(T_H + 459\right)\right]} \qquad (5)$$

$P_1$ and $P_2$ will of course be readily available from the altimeter readings, but in order to obtain the value of $\delta T$ it is advisable always to measure the lapse rate at the relevant altitude preferably before and after the experiment.

In order to decide whether $\delta T$ or $\delta P$ can be ignored equations (6) below can be used. If $\delta\alpha_T$ is the error in $\alpha$ due to $\delta T$ and $\delta\alpha_P$ is the error due to $\delta P$ when equation (3) is used then:-

$$\left.\begin{aligned} \frac{\delta\alpha_T}{\alpha} &\doteqdot \frac{\delta T}{T_H - T_L} \\ \frac{\delta\alpha_P}{\alpha} &\doteqdot \frac{\delta P}{P} \end{aligned}\right\} \qquad (6).$$

In practice $\delta\alpha_P$ is usually negligible and $\delta\alpha_T$ is small so that $P_1 = P_2 = P$ and $\delta T$ can be applied directly as a correction to $T_L$ which can be used in equation (3).

## The effect of random errors in temperatures, speed and height readings on the computed value of the speed correction coefficient

If the respective random errors in $T_L$, $T_H$, $V_H$, $V_L$ and P are $r_1$, $r_2$, $r_3$, $r_4$ and $r_5$ respectively, then there will be an error in $\alpha$ which is given by:-

$$r_\alpha^2 = \left(\frac{\partial\alpha}{\partial T_L}\right)^2 r_1^2 + \left(\frac{\partial\alpha}{\partial T_H}\right)^2 r_2^2 + \left(\frac{\partial\alpha}{\partial V_H}\right)^2 r_3^2 + \left(\frac{\partial\alpha}{\partial V_L}\right)^2 r_4^2 + \left(\frac{\partial\alpha}{\partial P}\right)^2 r_5^2 \qquad (7).$$

By performing the differentiations on equation (3) and taking typical values of the random errors it is possible to estimate the relative effect of these various random errors on $r_\alpha$ .

In the following chosen example:-

| Height | P | T | $V_H$ | $V_L$ | $r_1$ | $r_2$ | $r_3$ | $r_4$ | $r_5$ |
|---|---|---|---|---|---|---|---|---|---|
| feet | mb | $^{\circ}F$ | knots | knots | $^{\circ}F$ | $^{\circ}F$ | knots | knots | mb |
| 5000 | 843 | 41 | 180 | 120 | 0.2 | 0.2 | 1.5 | 1.5 | 0.5 |

$r_\alpha$ was computed to be approximately 0.15, the main contribution being due to temperature errors. Speed errors alone would cause 0.06 and altitude errors 0.002.

A similar example worked out for typical conditions at 21000 feet showed that compared to 5000 feet the effect of random temperature errors is nearly halved, that of pressure errors nearly doubled while that of speed errors was about the same. Hence successive determinations of $\alpha$ are likely to be more consistent at the higher levels and this in fact is shown in the results of Part I of this report.

It appears incidentally from a consideration of the components above that random errors in speed and height readings become more important in the high speed run than in the low speed run.

## The flying necessary to obtain a given accuracy in a speed correction coefficient

If a single determination of the speed correction coefficient has a probable error of say 0.15, the number of determinations required to obtain a probable error of 0.05 will be nine and of 0.01 two hundred and twenty five. This latter is prohibitive and 0.05 is probably the highest accuracy obtainable in practice for a calibrated aircraft.

## Discussion

In addition to the errors discussed in the preceding paragraphs other inconsistencies always appear in the practical measurement of temperature from aircraft. Turbulence may affect the readings obtained from a pitot static system (4) and also the performance of the thermometer itself (M.R.P. No. 677 and 744). A "pitot position error" may also be produced due to small angles of pitch and yaw. In general an apparently random error of about 0.3°F usually appears in all attempts to measure temperature, e.g. in sequences of the same average temperature over say 5 minute runs, and this may be due to any or all of the above causes rather than a small change in the mean air temperature.

It appears therefore that it is not likely that the air temperature measured even from a research aircraft will have an accuracy better than about 0.5°F. In view however of the inhomogeneity of the atmosphere this is usually satisfactory for most practical purposes. However in a fast aircraft which has not been specially calibrated a single determination of air temperature may easily be up to 5°F in error.

It is considered that any further experimental work on this subject should now await the availability of fast aircraft with a wide speed range at great heights. It may become possible with this aircraft to detect with certainty whether the speed correction coefficients of these aircraft thermometers do in fact vary with height as the importance of the random errors should then decrease. The considerations outlined above should be taken into full account when this work is done, and in particular great attention should be paid to position error effects as these increase considerably with speed and with altitude.

References

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Taking a mean value of $V_I$ = [illegible] kts. the values of Re above come out as approx. Re = $10^5$. In the circumstances therefore it is not expected that the tests described would be likely to show up any variations due to boundary layer phenomena.

The variations in the $\alpha$ of the first flat plate are within the limits of definition claimed on p.13 of the report. The increase of $\alpha$ in the second flat plate may be due to bad siting as explained in MRP. 677. The increase in $\alpha$ of the M.O. conical-head thermometer is in accordance with theory and follows from the existence of a velocity gradient along the cone. Steps have been taken on later models to reduce the effects of this gradient. The variations in $\alpha$ of the Sangamo-Weston thermometer are believed to arise from insufficient thermal insulation between the sensitive element and the nose of the thermometer. The most astounding result of all is the effect of the trip wires on the $\alpha$ of the flat plate. The explanation given in the paper is the most plausible and agrees with observations recorded by Ackeret (Ric. Sci. 20, Dec. 1950, 1926-7). No explanation other than structural damage can be given for the shift in the value of $\alpha$ of the first flat plate between the two series of flights.

The suggestion on p.3 to direct attention to reducing the coefficient is good practice where approximate temperatures are wanted but, in an aircraft whose specific purpose is to obtain accurate temperature observations, it is not thought that a thermometer carrying devices to reduce the $\alpha$ can be relied upon.

On p.8 5(b) the remarks about the position-error correction apply only if the methods of Charnley and Fleming are used. No correction for compressibility need be applied to the position-error if the methods laid down in MRP. 604 are followed.

(Sgd.) D.D. CLARK.

M.O.17
4/5/53.

Control Number. HD-31723

(Control Number AD-37923)
C.P. 8053.

PROVISIONAL

M.R.P. 823 REVISED

S.C. II/156

4th March, 1954

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 823 (Revised) by Mr. A.F. Jenkinson A copy is available in the Library of the Meteorological Office."

AD37923

# AIR MINISTRY
# METEOROLOGICAL RESEARCH COMMITTEE

## The average vector wind distribution of the upper air in temperate and tropical latitudes

by

A.F. Jenkinson (Meteorological Office)

### Summary

Charts have been drawn showing, for the months of January, April, July and October, and for the area between the latitudes of 60°N. and 50°S., the isotachs of wind speed and the streamlines of wind flow at the levels of 500, 300, 200 and 100 mb.

Note

(Publication of [illegible] prepared in the Upper Air Branch of the Meteorological Office, of mean air temperatures over the world at the levels of 700, 500, 300, 200 and 100 mb. for the months of January, April, July and October [illegible] drawn at intervals of 2°C.

These have been revised recently in some areas (India, Australia and South Africa) to include the latest available data.)

Theory

If $s(p_1p_2)$ is the thermal vector wind between two pressures $p_1$ and $p_2$ and $n$ is a unit vector directed vertically, then

$$\lambda s(p_1p_2) = n \times \text{grad } h(1,2) \qquad (1)$$

$$\text{where } \lambda = 2\Omega \sin\phi$$

$$h(1,2) = \int_{p_1}^{p_2} (RT/g)\, d(\log p)$$

$\Omega$ is the earth's angular velocity

$R$ is the gas constant

$g$ is the acceleration due to gravity

$T$ is the absolute temperature

$$h(1,2) = (R/g)\, T(1,2) \log_e (p_1/p_2) \qquad (2)$$

where $T(1,2)$ is the mean value of $T$ on a $T$, $\log p$ diagram between the pressures $p_1$ and $p_2$

$$s(p_1p_2) = k(1,2) \left[ n \times \text{grad } T(1,2) \right] \qquad (3)$$

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AD37923

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

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## Summary

Charts have been drawn showing, for the months of January, April, July and October, and for the area between the latitudes of 60°N. and 50°S., the isotachs of wind speed and the streamlines of wind flow at the levels of 500, 300, 200 and 100 mb.

## Data

Preliminary charts have been prepared in the Upper Air Branch of the Meteorological Office of upper air temperatures over the world at the levels of 700, 500, 300, 200, 150 and 100 mb. for the months of January, April, July and October, with isotherms drawn at intervals of 2°C.

These were then amended slightly in some areas (India, Australia and South Africa) to include the latest available data.[1]

## Theory

If $\theta\ (p_1 p_2)$ is the thermal vector wind between two pressures $p_1$ and $p_2$, and n is a unit vector directed vertically, then

$$\lambda\, \theta\ (p_1 p_2) = n \times \text{grad}\ h\ (1,2) \qquad (1)$$

$$\text{where} \quad \lambda = 2\, \Omega \sin \phi$$

$$h(1,2) = \int_{p_1}^{p_2} (RT/g)\ d(\log p)$$

$\Omega$ is the earth's angular velocity

R is the gas constant

g is the acceleration due to gravity

T is the absolute temperature

$$h(1,2) = (R/g)\ T(1,2) \log_e\ (p_1/p_2) \qquad (2)$$

where $T(1,2)$ is the mean value of T on a T, log p diagram between the pressures $p_1$ and $p_2$

$$\theta\ (p_1 p_2) = k(1,2) \left[ n \times \text{grad}\ T(1,2) \right] \qquad (3)$$

$$\text{where} \quad k\ (1,2) = (R/\ \lambda\ g) \log_e\ (p_1/p_2)$$

If the average monthly temperature at various pressure levels p is plotted against p the curve is essentially two straight lines, with a discontinuity at the tropopause.

/The

G.366828/JWP/6/53/40.

The layers chosen for the computation of thermal vector winds were 1000 - 500, 500 - 300, 300 - 200, 200 - 150, 150 - 100 mb. Apart from the layer including the tropopause the average curve of T, log p differs most from a straight line in the layer 1000 - 500 mb.; but even in this layer the mean temperature T (1000,500) is very nearly T (707) the temperature at 707 mb. ($\log 707 = \frac{1}{2} \left[\log 1000 + \log 500\right]$). At neighbouring places the way in which the average temperature curve differs from a straight line is nearly identical, and the rate of change from place to place of the difference between T(1000,500) and T(707) is of a very much smaller order than the rate of change of T(707), which is very nearly the rate of change of T(700).

Hence, very approximately

$$\theta\,(1000,500) = k\,(1000,500) \left[ n \times \operatorname{grad} T(700) \right] \qquad (4)$$

That is, the thermal wind of the layer 1000 - 500 mb. is in the direction of the 700 mb. isotherms with magnitude proportional to the gradient of temperature between the isotherms.

For the other layers above 500 mb. not including the tropopause, the mean temperature between any two levels is very nearly half the sum of the temperatures at the two levels; and the gradient of the field of mean temperature is very approximately equal to half the vector sum of the gradients of the temperature fields of the two levels.

For the layer including the tropopause, the curve of T, log p is two straight lines. Curves for neighbouring places are similar and the rate of change from place to place of the difference between T(1,2) and $\frac{1}{2}\left[T(p_1) + T(p_2)\right]$ is very much less than the rate of change of $\frac{1}{2}\left[T(p_1) + T(p_2)\right]$. Hence also for the layer including the tropopause the gradient of the field of mean temperature is very approximately equal to half the vector sum of the gradients of the temperature fields of the two levels.

Thus the thermal wind of the layer 500-300 mb. is equal to half the sum of two vectors which are respectively directed along the isotherms of the 500 and 300 mb. temperature fields and with magnitudes proportional to the gradients of temperature at those levels; and similarly for other layers.

In the notation of equation (4),

$$\theta\,(500,\ 300) = \tfrac{1}{2}\, k\,(500,\ 300) \left\{ \left[ n \times \operatorname{grad} T\,(500) \right] + \left[ n \times \operatorname{grad} T\,(300) \right] \right\}$$

..................................................................................................................

$$\theta\,(150,\ 100) = \tfrac{1}{2}\, k\,(150,100) \left\{ \left[ n \times \operatorname{grad} T\,(150) \right] + \left[ n \times \operatorname{grad} T\,(100) \right] \right\} \qquad - (5)$$

## Method

A fixed grid of 234 points was chosen, the intersection of longitudes 180°W. (20°) 180°E. and latitudes 20°N. (5°) 40°N., 50°N., 60°N. and latitudes 20°S. (5°) 40°S., 50°S.

For each of the months January, April, July and October, at each grid point for the levels of 700, 500, 300, 200, 150 and 100 mb. measurements were made of the vector quantities whose directions were along the isotherms and magnitudes equal to the gradients of the temperature field. From these, substituting the appropriate values of the constants k, the thermal winds were obtained for the layers 1000-500, 500-300, 300-200, 200-100 mb., with speeds measured to the nearest knot and directions to the nearest five degrees.

The "surface" geostrophic vector wind at each of the 234 grid points was measured from mean sea level pressure charts, with isobars drawn at 4 mb. intervals, prepared in the World Climatology Branch of the Meteorological Office. Over the oceans a check was obtained from the charts of average winds for five degree squares prepared in the Marine Branch of the Meteorological Office[23]; and for some doubtful points over the continents from available manuscript data in the Meteorological Office.

The vector winds at 500, 300, 200 and 100 mb. were obtained by successive vector addition of the "surface" and thermal vector winds.

The vectors were plotted on sixteen Mercator charts which are reproduced as Figs. 1 to 16[x]. Isotachs of wind speed in kt. (full lines) and stream lines (pecked lines with direction arrows) were drawn. Most of the stream lines were drawn as isopleths of approximate contour height in hundreds of geopotential ft. The contour heights have been entered for the benefit of possible users but it should be understood that they are not complete in themselves. Areas with resultant wind speed less than 10 kt. were hatched in the tropics.

To help in drawing the charts, wind data were worked up for a number of stations, chiefly for three year periods between 1949 and 1952; and with the help of these data and other available data it was considered that the charts could be drawn up even in the tropical regions with considerable accuracy. Lists are appended of data worked up, data recently published or made available, and also of older material examined.

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Footnote

[x] A limited number of charts, approximately double the size of those in the report, are also available.

## APPENDIX I

Wind data worked up for three-year period

Athens [2,3]; Hanover[4]; Cairo[7].

San Juan (all months) and Bismarck, Columbia, Greensboro, Little Rock, Oakland, San Antonio, Tatoosh, Caribou, Oklahoma, St. Cloud, Big Spring, Brownsville, Charleston, Miami, New Orleans for the month of July only.[5]

Canton I, Honolulu, Midway I, Albrook Field, Marshall I, Guam, Manila, Wake I, Rome[3].

Gibraltar, Benina, Malta, Nicosia, Nairobi, Habbaniya, Bahrein and Aden from manuscript data in the Meteorological Office.

Lagos (1952 only)

Japanese stations, chiefly Wakkanai, Tateno, Shionomisaki, (1950-1).[6]

## APPENDIX II

Wind data recently published or made available

Larkhill, Lerwick[8]; Poona (July 1949-50)[9];
Auckland, Nandi[10]; Singapore[11]; Hong Kong[12].

Falkland I., Aldergrove as manuscript data available in the Meteorological Office.

Mean winds for Japanese stations for the months of January, April and July 1952[6].

## APPENDIX III

Older material examined

Cirrus cloud observations, Napier Shaw [13].
Lindenberg[14]; Batavia[15]; Mauritius[16]; Apia[17];
Ghanzi[18].
Cuyaba, Curityba, Santiago, Meteor[19].
Pilot balloon data for India [20,21]

Vector winds computed from medium and cirrus cloud observations for the United States and the Caribbean [22].

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21. Venkiteshwaran, S.P. Winds at 10 kms. and above over India and its neighbourhood. Mem. Ind. Met. Dept. Vol.XXVIII, Part II. Delhi, 1950.

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9. Venkiteshwaran, S.P. and S. Yogannarayanan. Radar measurements of upper winds over Poona during the southwest monsoon. Indian Journal of Meteorology and Geophysics, Vol.2. No.3. Delhi, July 1951.

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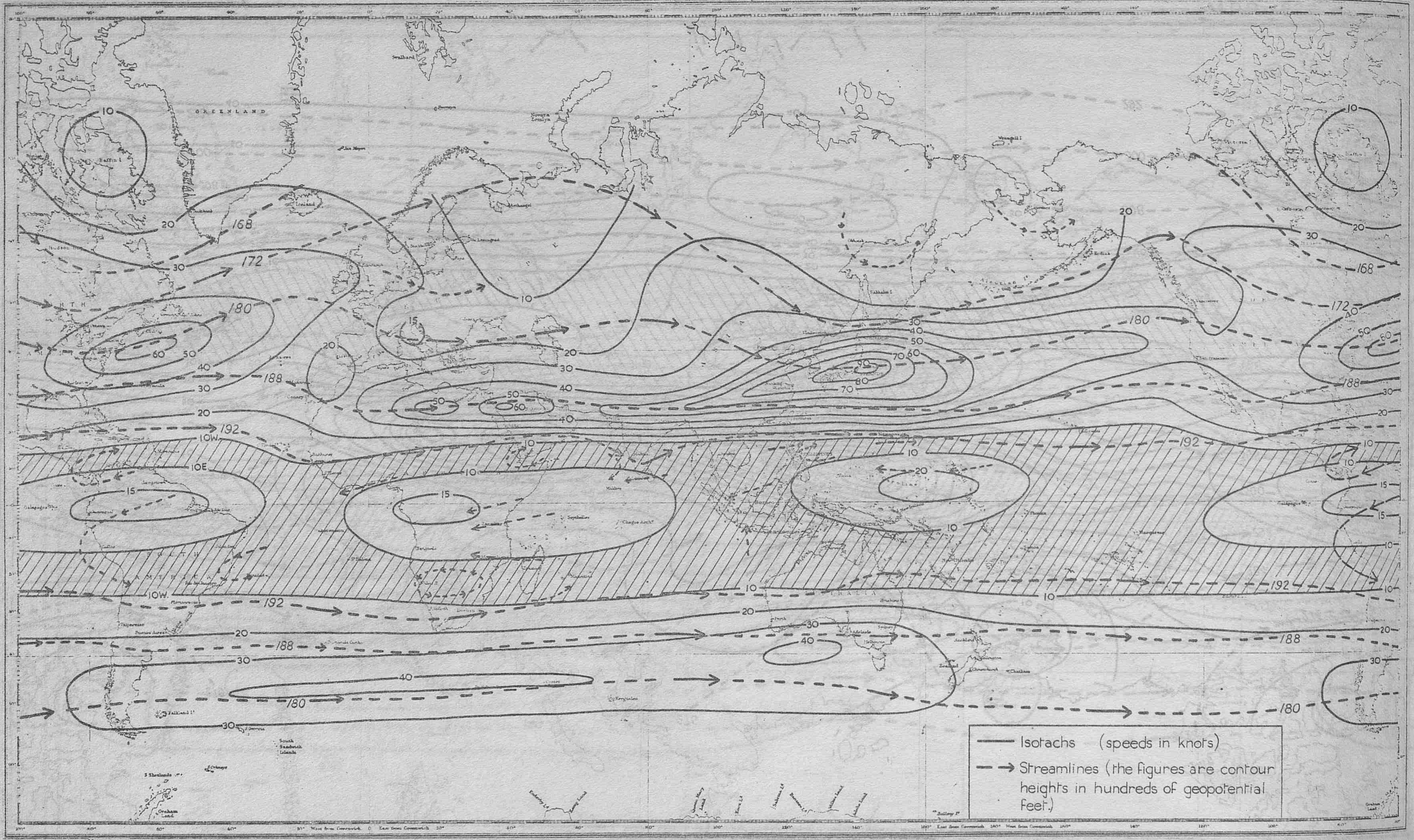
JANUARY – 500mb.

PROVISIONAL

# Streamlines and isotachs of the average vector wind distribution

M.R.P. 823



GREENLAND
Svalbard
Novaya Zemlya
Iceland
Baffin I.
Hudson Bay
Falkland I.
South Sandwich Islands
Kerguelen
Galapagos
Seychelles
Chagos Arch.
Maldives
10
20
30
40
50
60
70
80
90
15
10W
10E
168
172
180
188
192
Isotachs (speeds in knots)
Streamlines (the figures are contour heights in hundreds of geopotential feet.)

FIG. 1

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JANUARY — 300mb.

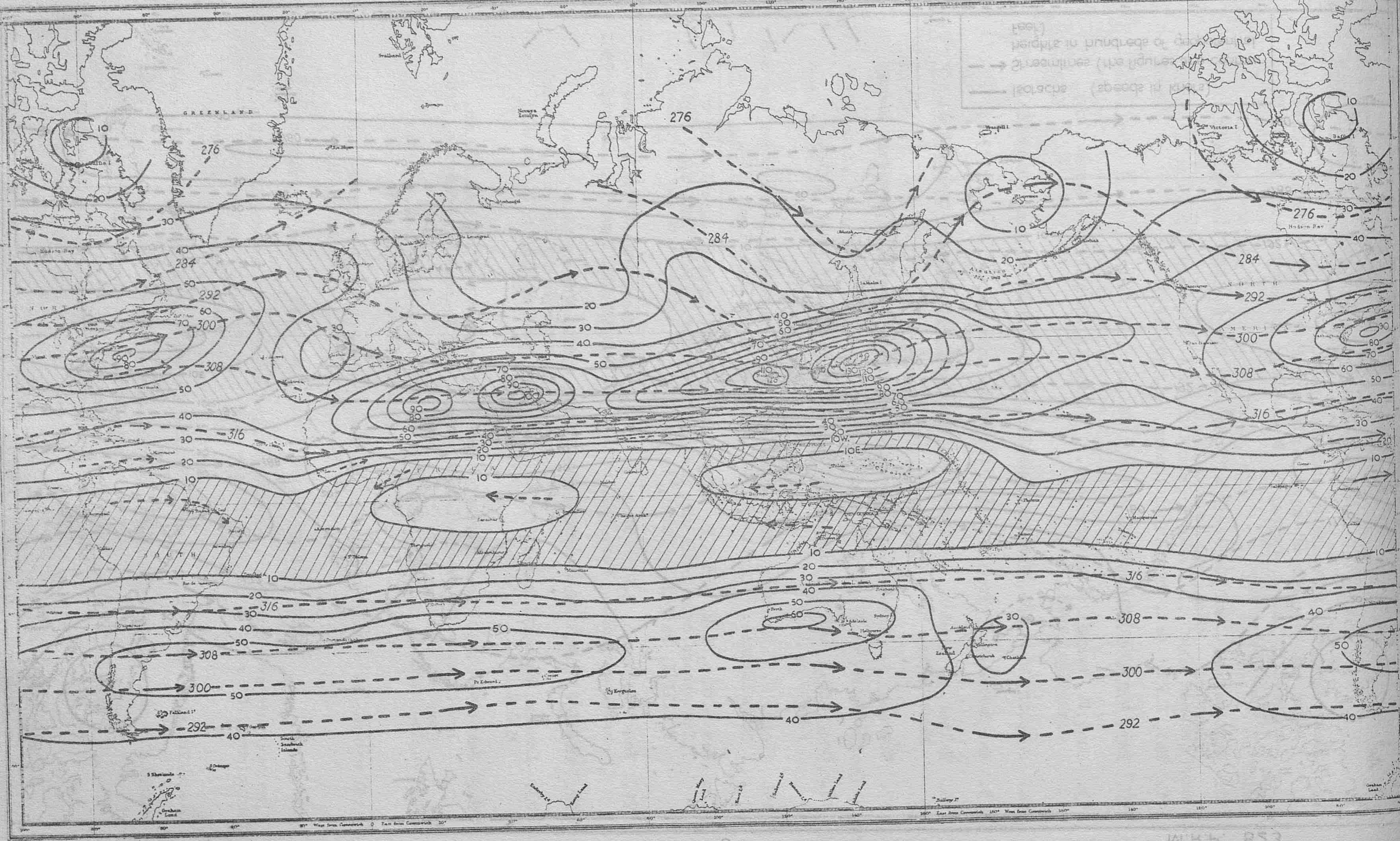
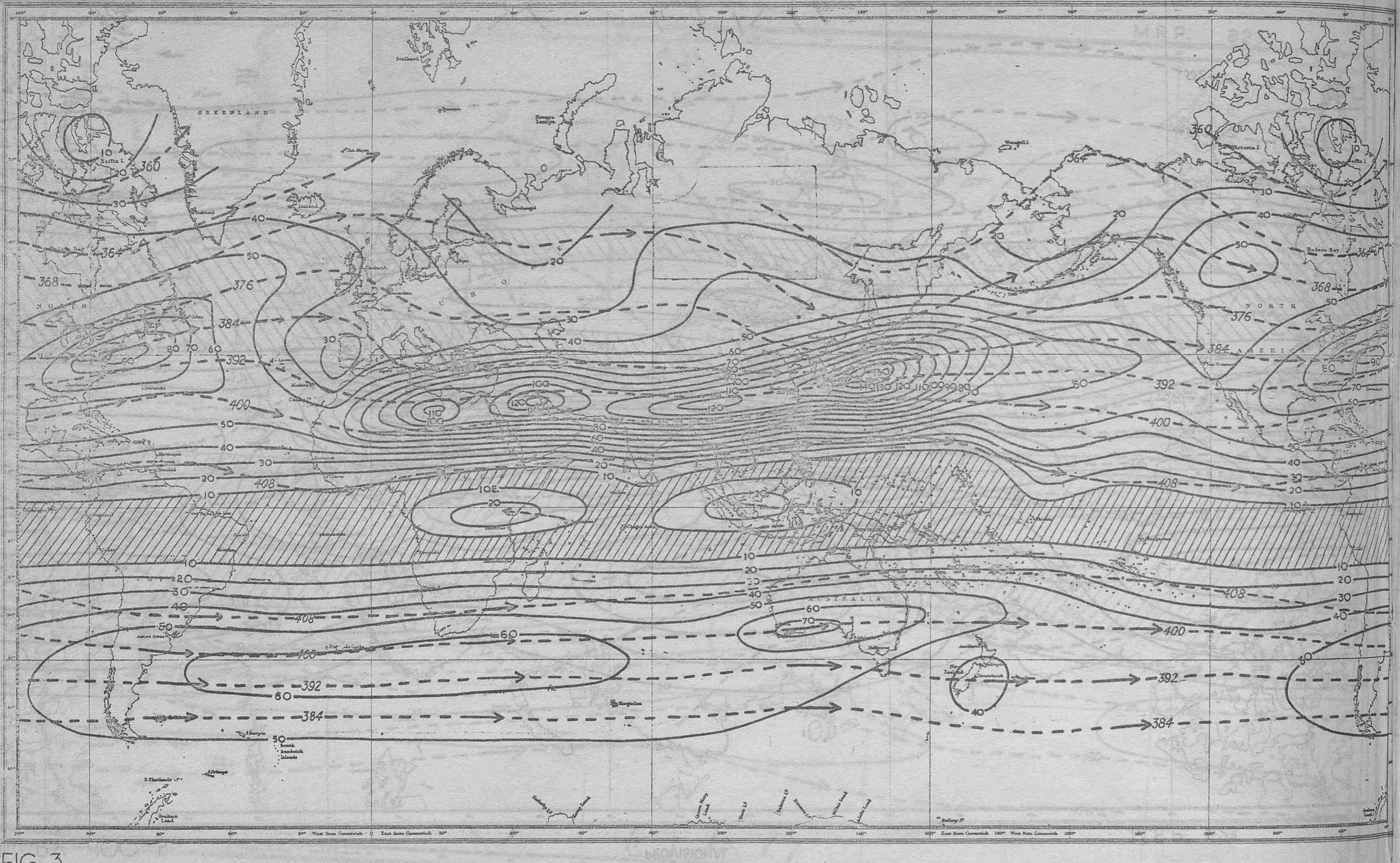

GREENLAND
276
284
292
300
308
316
10
20
30
40
50
60
70
80
90
110
120
130
140
10W
10E
NORTH
SOUTH
Hudson Bay
Novaya Zemlya
Svalbard
Falkland I.
Kerguelen
Victoria I.
Baffin I.
Wrangel I.
Aleutian Is.
South Sandwich Islands
S. Orkneys
S. Shetlands
Graham Land
Pr. Edward I.
Crozet
Perth
Adelaide
Sydney
Melbourne
Chatham
Mauritius
Madagascar
Rio de Janeiro
West from Greenwich
East from Greenwich

FIG. 2

JANUARY – 200 mb.

PROVISIONAL

M.R.P. 823



GREENLAND
Iceland
Svalbard
Novaya Zemlya
Baffin I.
Victoria I.
Hudson Bay
NORTH
AMERICA
EURO
AUSTRALIA
New Zealand
Kerguelen
South Sandwich Islands
360
364
368
376
384
392
400
408
10
20
30
40
50
60
70
80
90
100
110
120
130
140
10E
West from Greenwich
East from Greenwich

FIG. 3

# JANUARY — 100 mb.

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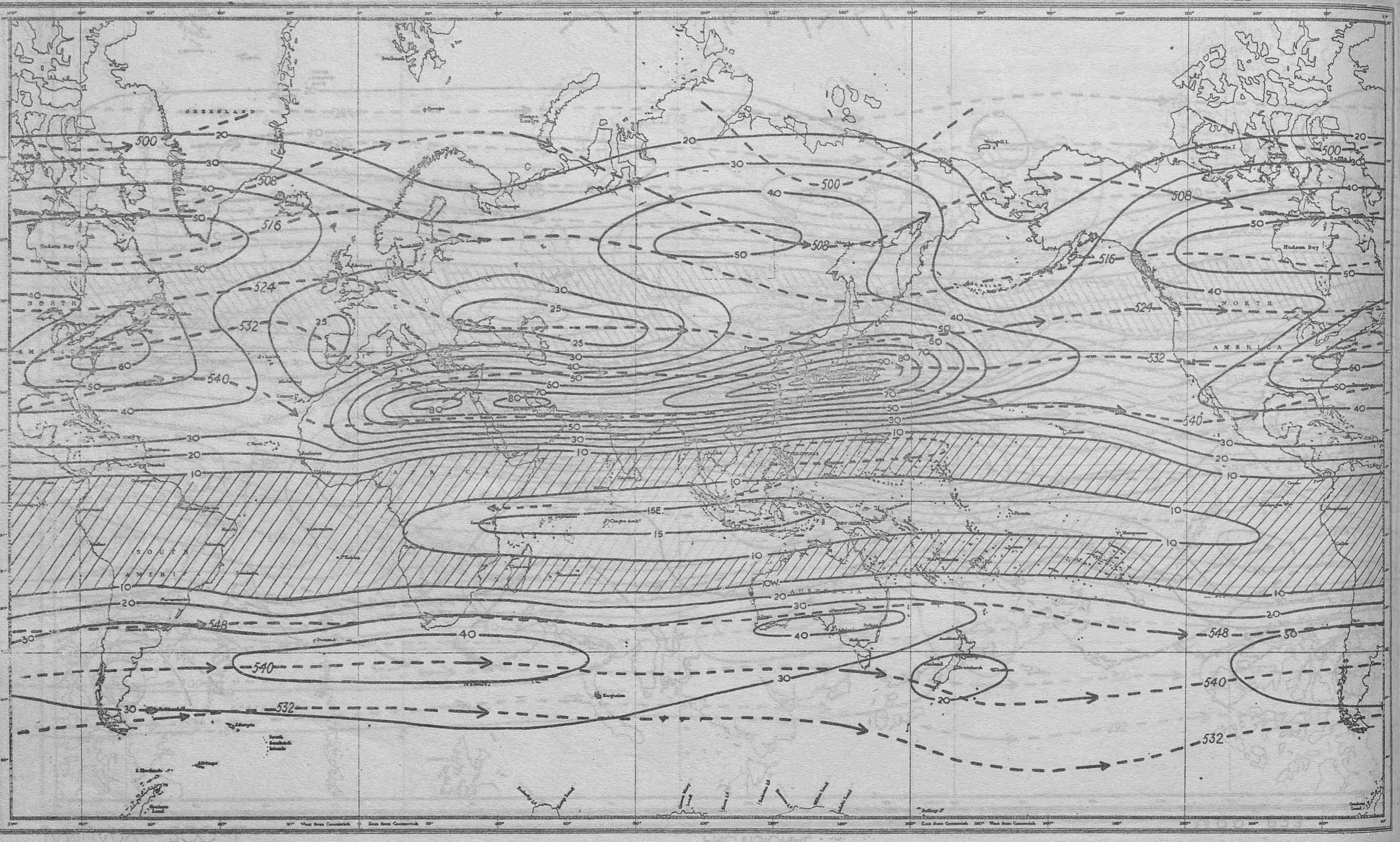
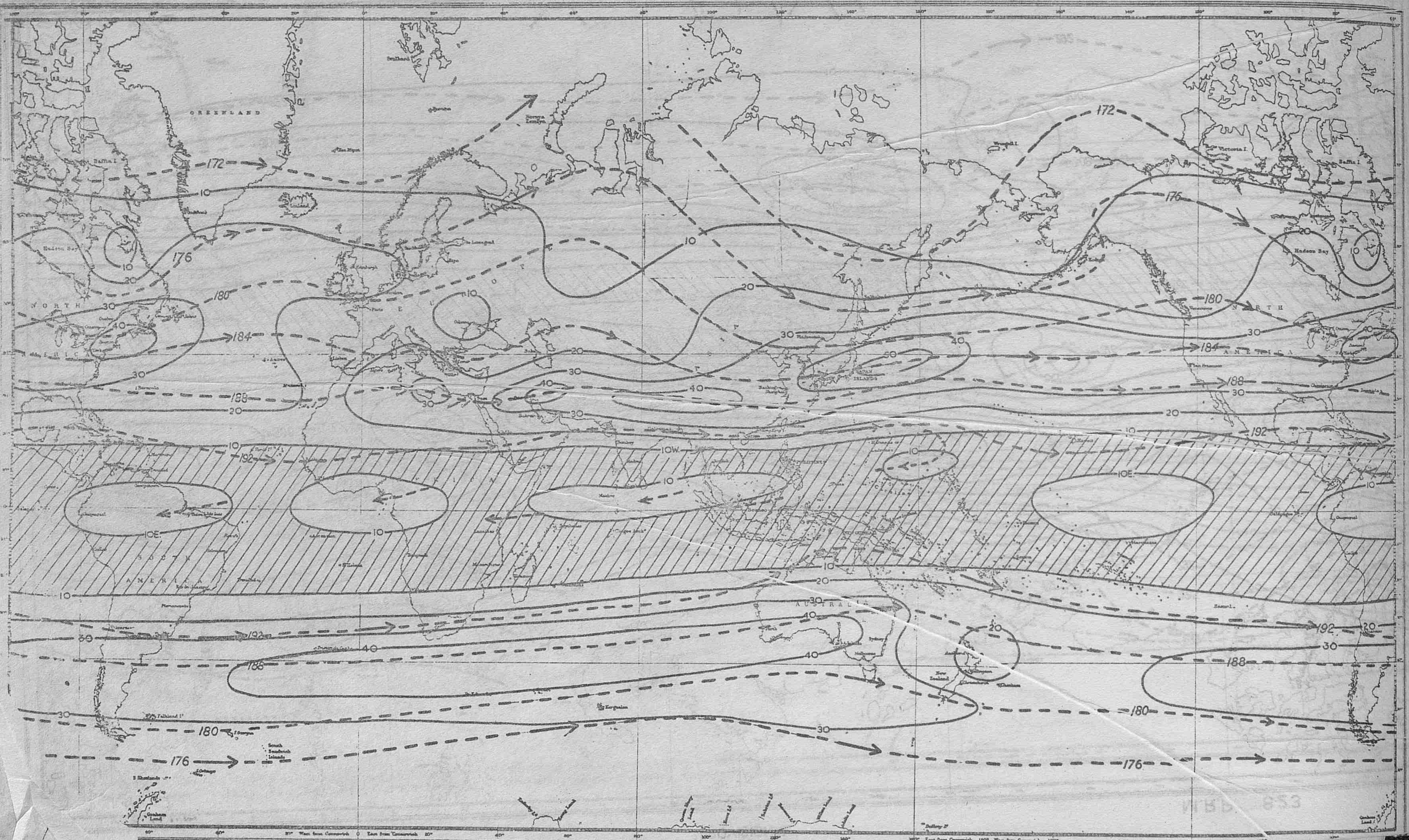


GREENLAND
Hudson Bay
NORTH AMERICA
SOUTH AMERICA
EUROPE
AFRICA
AUSTRALIA
500
508
516
524
532
540
548
540
532
20
30
40
50
60
70
80
90
100
25
10
15E
15
10W

FIG. 4

APRIL – 500mb. PROVISIONAL M.R.P. 823

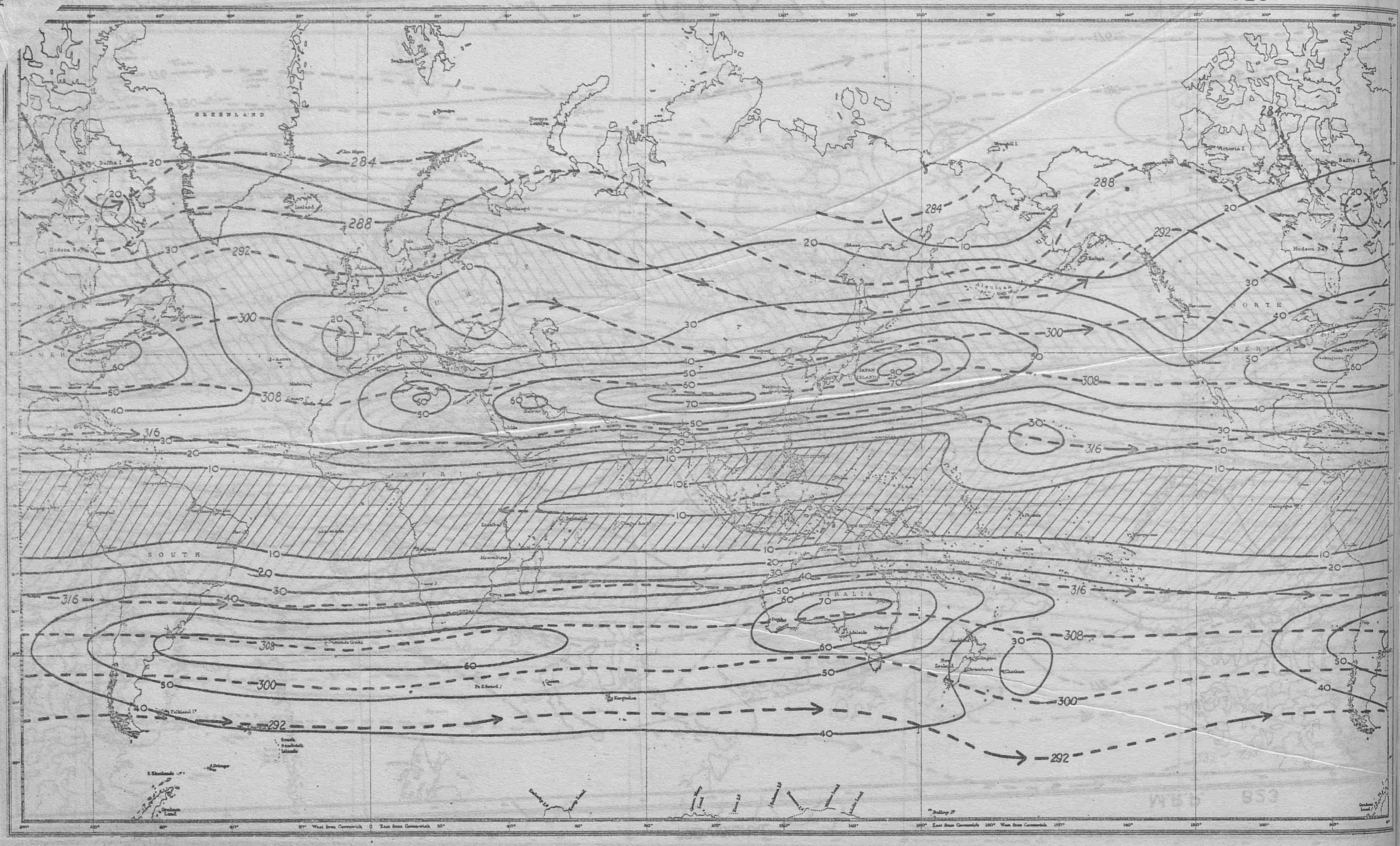


GREENLAND
Hudson Bay
NORTH AMERICA
SOUTH AMERICA
AUSTRALIA
New Zealand
172
176
180
184
188
192
10
20
30
40
50
10E
10W

-300mb.

PROVISIONAL

M.R.P. 823



GREENLAND
EUROPE
AFRICA
AUSTRALIA
NORTH AMERICA
SOUTH
JAPAN ISLANDS
Borneo
Iceland
Hudson Bay
Victoria I
Falkland Is
New Zealand
284
288
292
300
308
3/6
10
20
30
40
50
60
70
80
10E

FIG. 6

APRIL - 200mb.

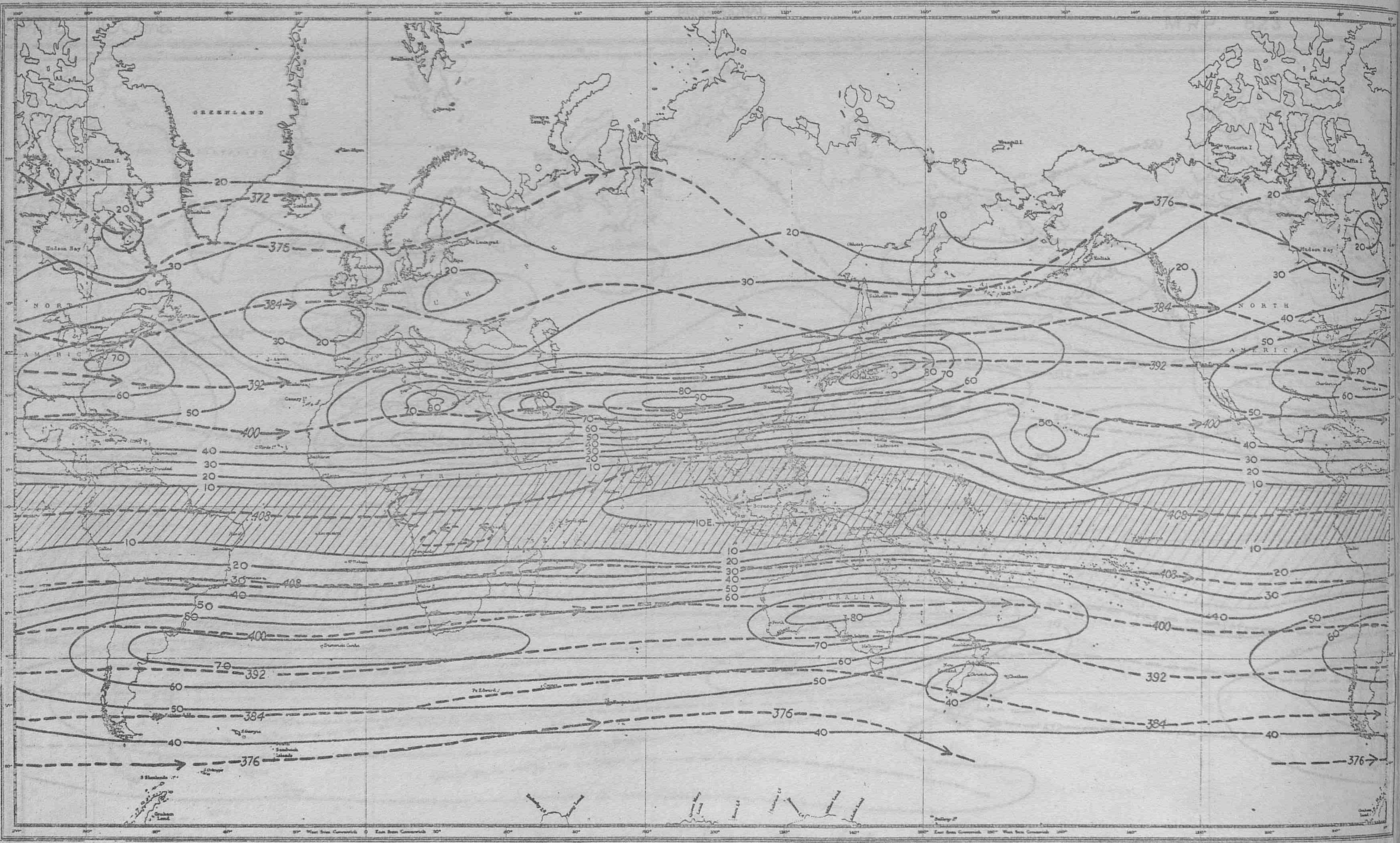
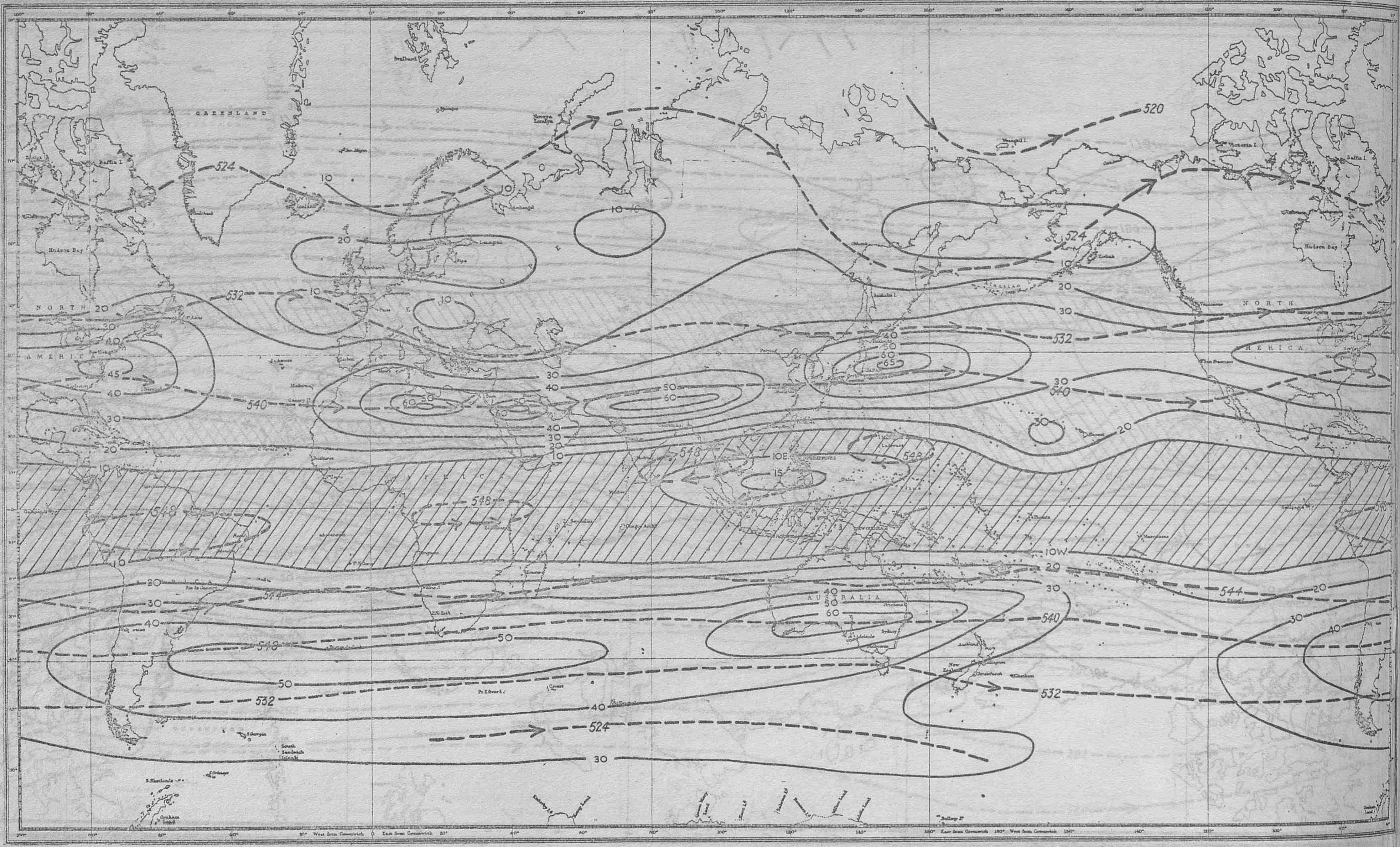

GREENLAND
NORTH AMERICA
AFRICA
AUSTRALIA
Hudson Bay
Baffin I.
Victoria I.
Iceland
Borneo
New Zealand
10E.
10
20
30
40
50
60
70
80
90
372
376
384
392
400
408

FIG. 7

APRIL - 100 mb.

PROVISIONAL

M.R.P. 823



GREENLAND
NORTH AMERICA
AUSTRALIA
Hudson Bay
524
520
532
540
544
548
10
20
30
40
45
50
60
65
10E
15
10W

FIG. 8

JULY – 500 mb.

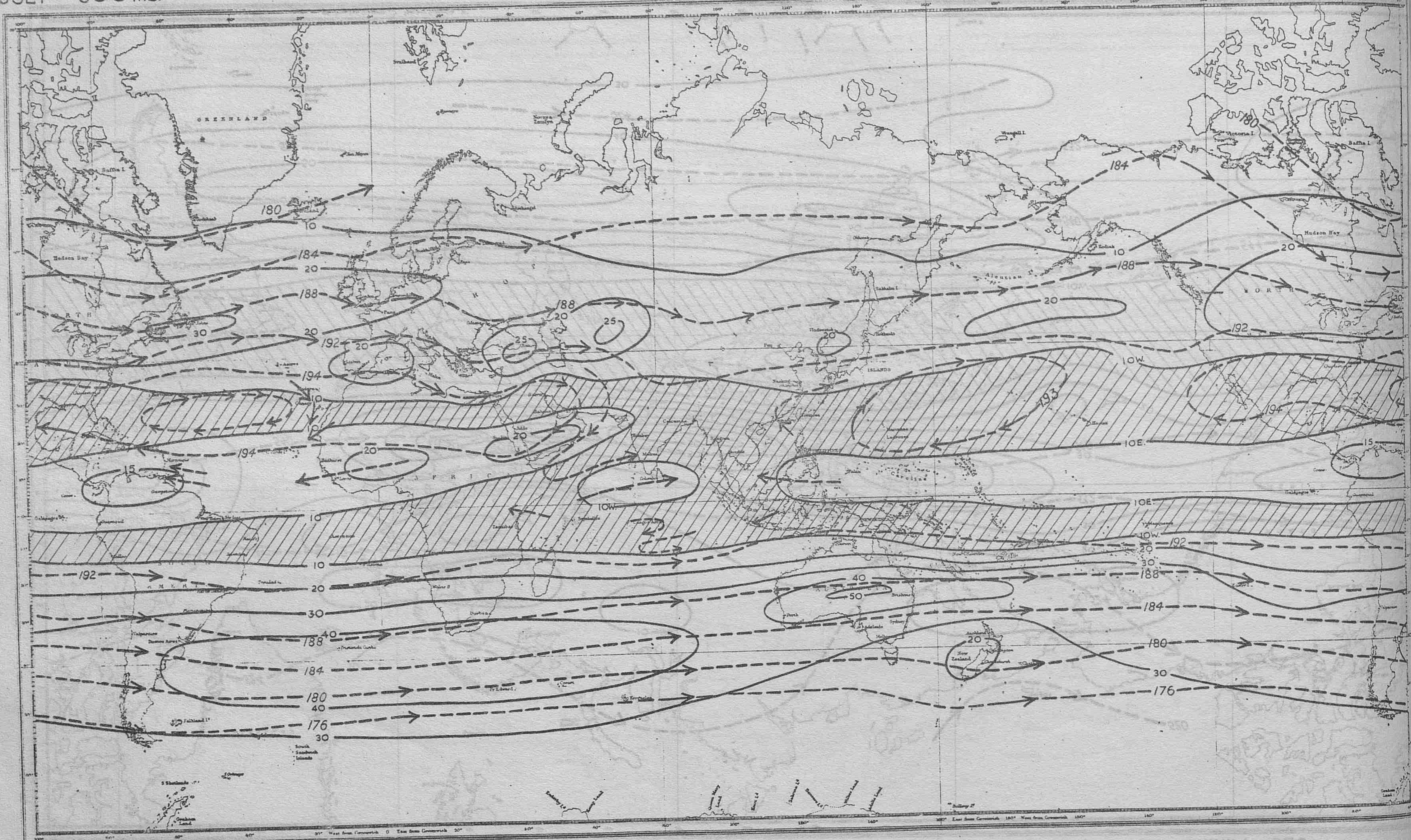
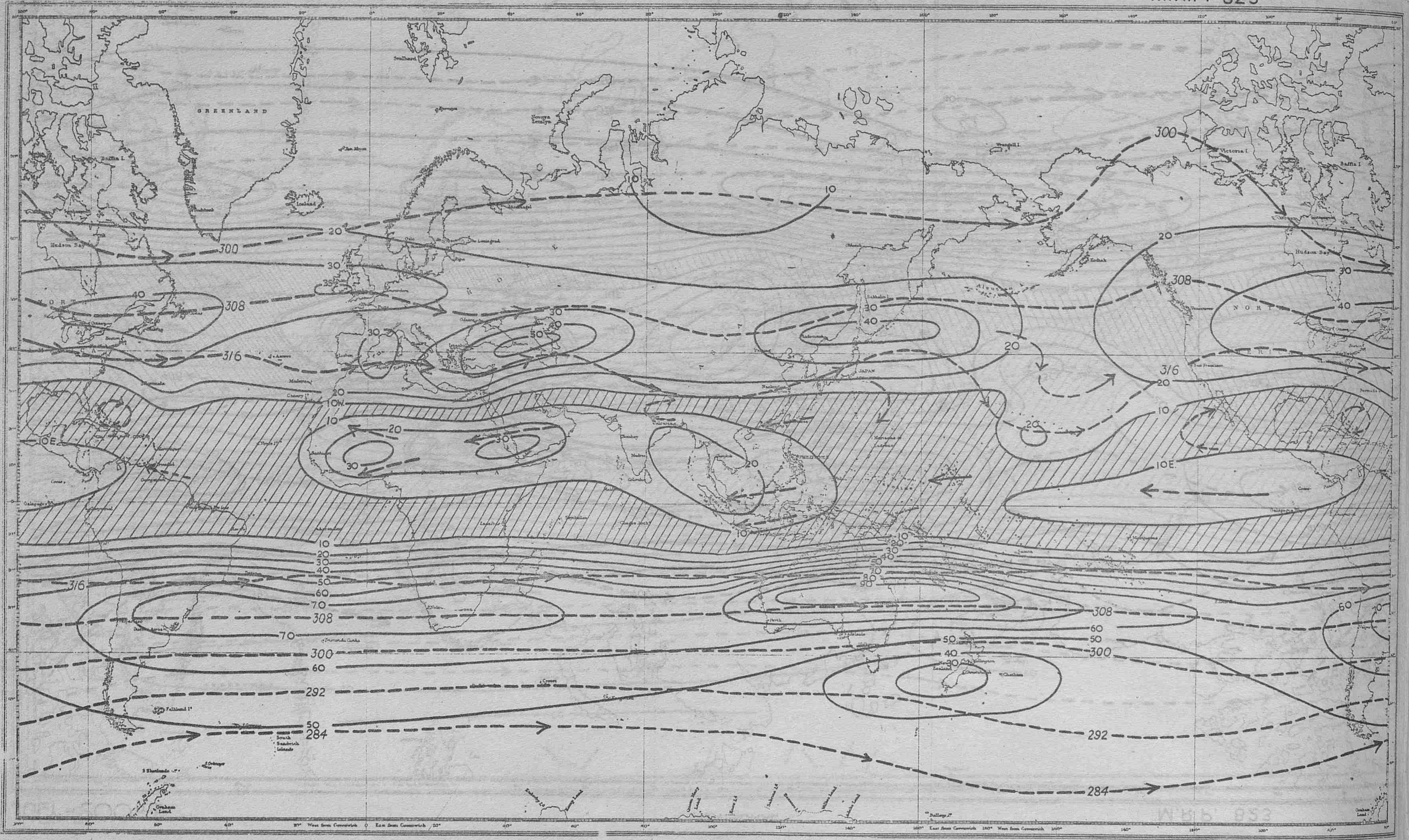

GREENLAND
Hudson Bay
Baffin I.
Victoria I.
Svalbard
Novaya Zemlya
Wrangel I.
Aleutian Is.
NORTH
AFRICA
AMERICA
AUSTRALIA
New Zealand
Falkland Is.
South Sandwich Islands
S. Shetlands
Graham Land
Kerguelen
Galapagos Is.
180
184
188
192
193
194
176
10
15
20
25
30
40
50
10W.
10E.

FIG. 9

JULY – 300mb.

PROVISIONAL.

M.R.P. 823



GREENLAND
Iceland
Svalbard
Novaya Zemlya
Hudson Bay
Baffin I.
Victoria I.
JAPAN
Wrangel I.
10
20
30
40
50
60
70
80
90
10W.
10E.
284
292
300
308
316

FIG. 10

JULY - 200 mb.

PROVISIONAL

M.R.P. 823

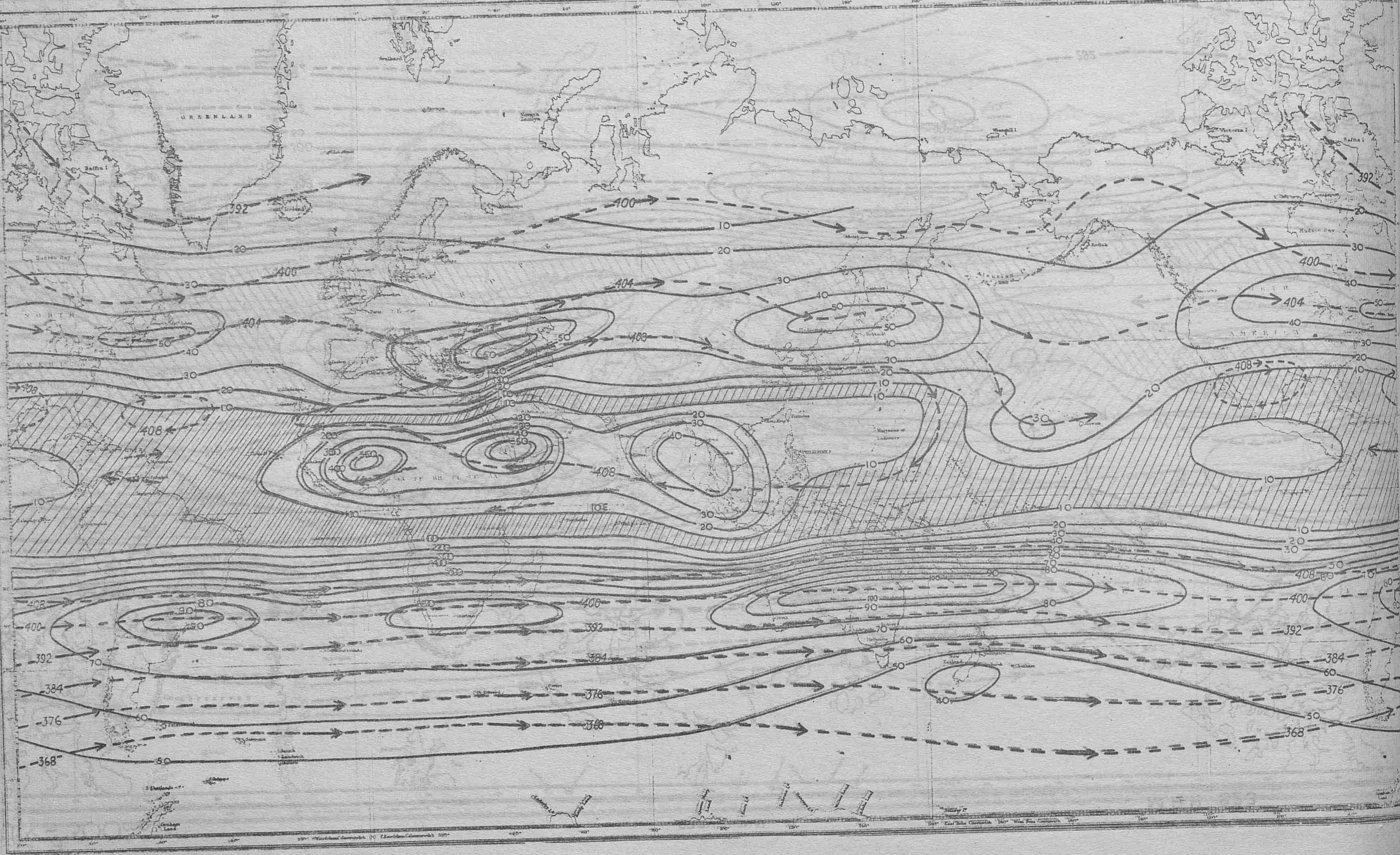

GREENLAND
392
400
404
408
400
392
384
376
368
10
20
30
40
50
60
70
80
90
100
IOE

FIG. 11

JULY - 100 mb.

PROVISIONAL

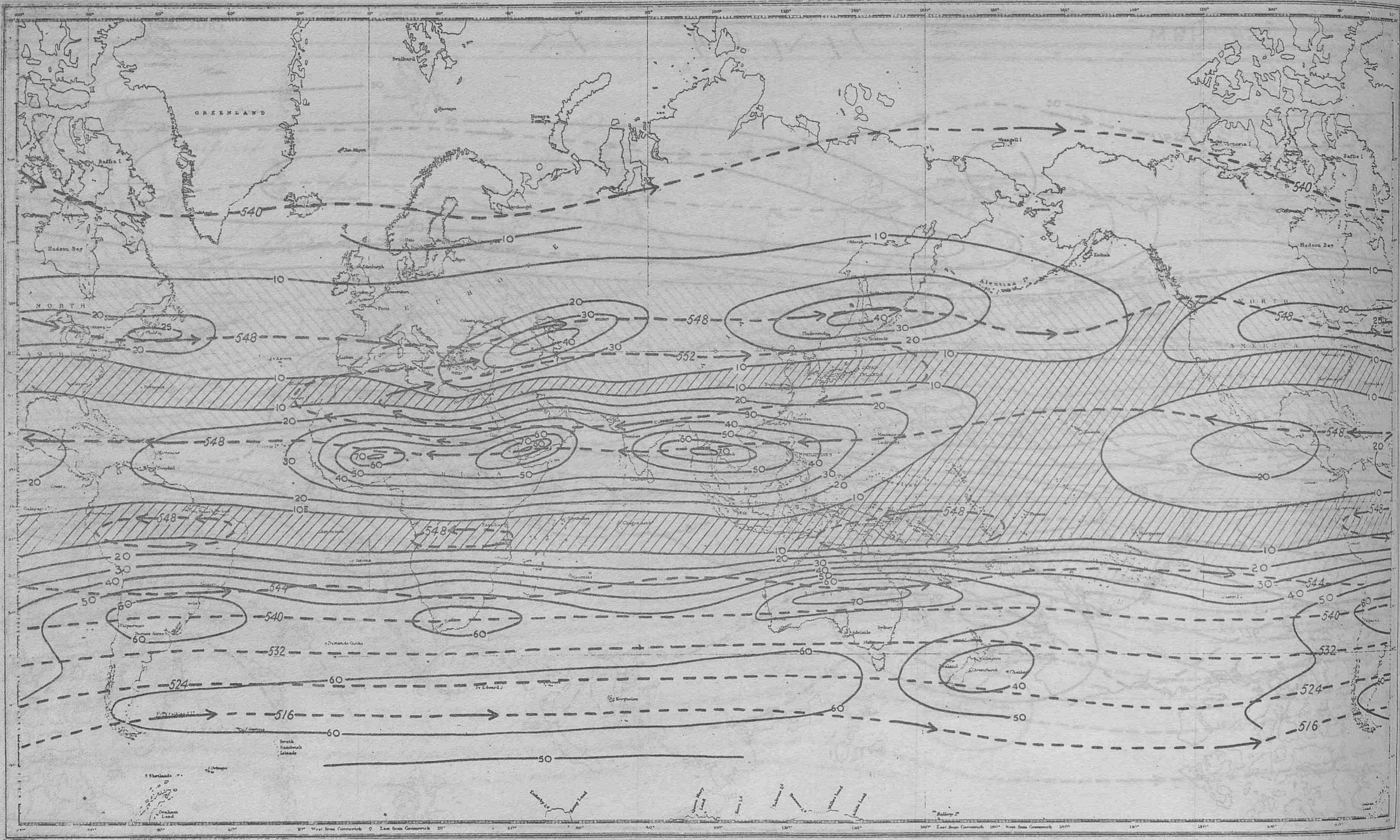

GREENLAND
Hudson Bay
Baffin I
Svalbard
Novaya Zemlya
Wrangel I
Victoria I
NORTH
AMERICA
EUROPE
AFRICA
PHILIPPINES
Aleutian Is
South Sandwich Islands
S Shetlands
Graham Land
Kerguelen
540
548
552
544
532
524
516
10
10E
20
25
30
40
50
60
70
West from Greenwich
East from Greenwich

FIG. 12

OCTOBER – 500mb.

PROVISIONAL

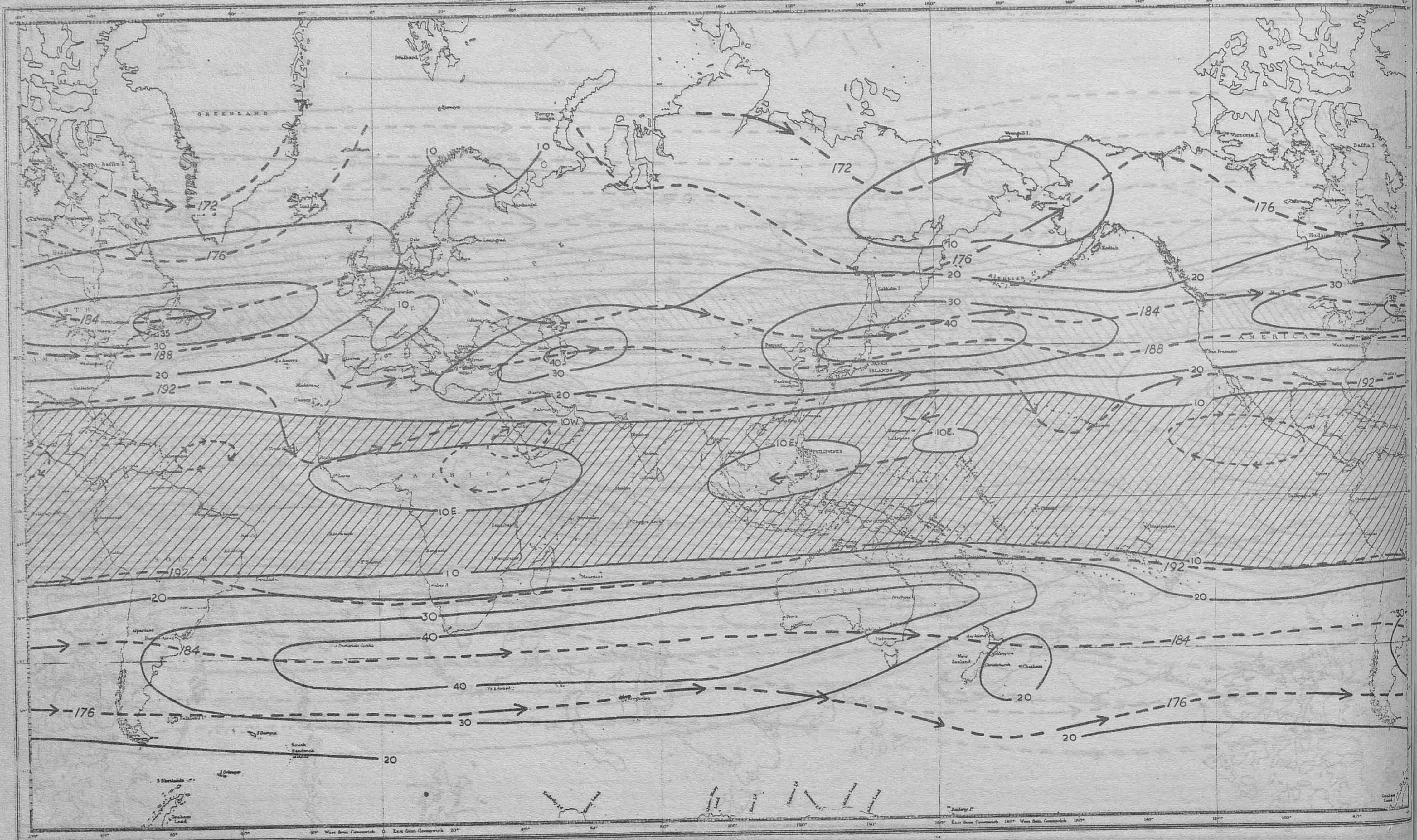
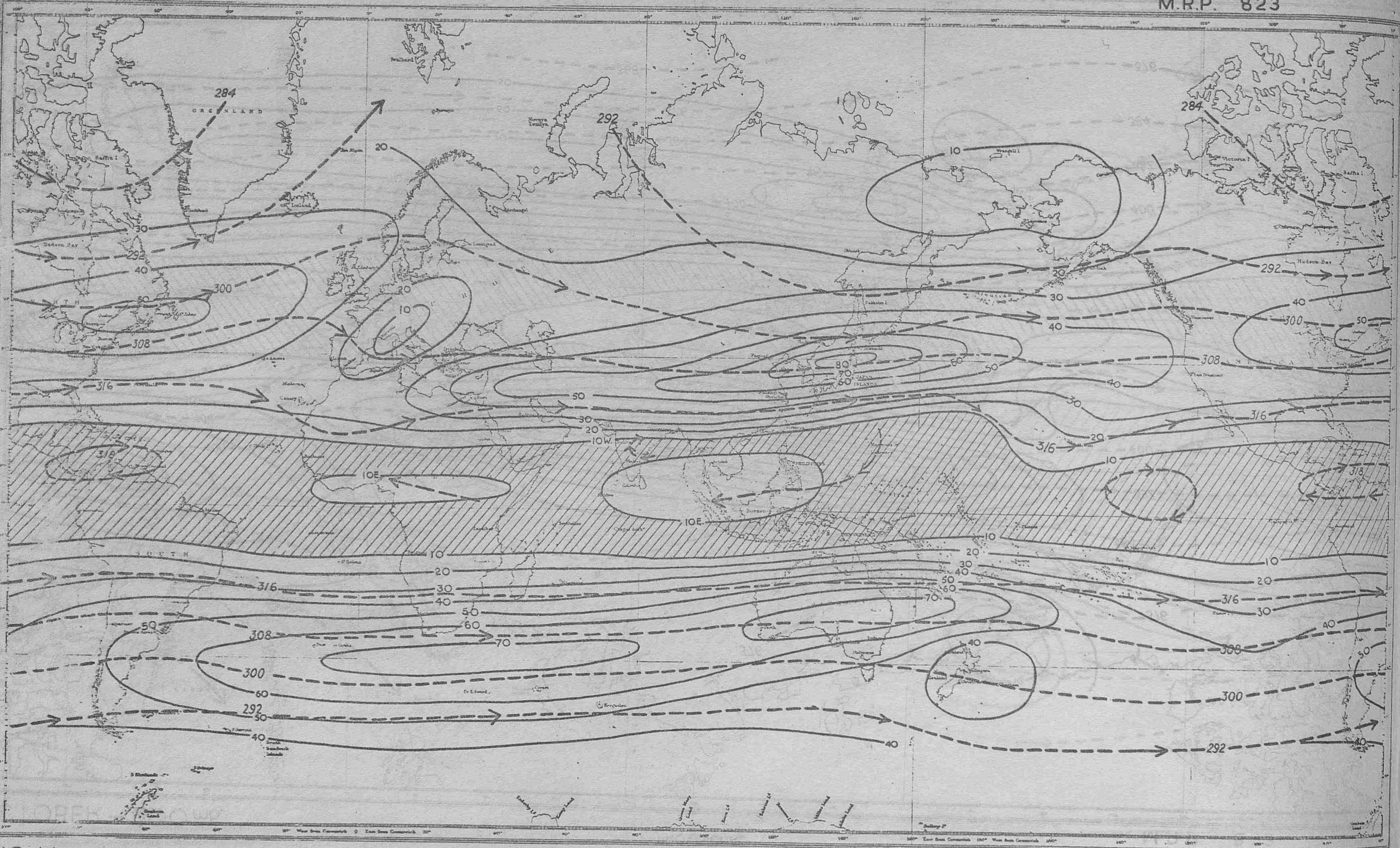

GREENLAND
AFRICA
AMERICA
AUSTRALIA
JAPAN ISLANDS
PHILIPPINES
New Zealand
172
176
184
188
192
10
20
30
35
40
10E.
10W.

FIG. 13

# OCTOBER — 300mb.

PROVISIONAL

M.R.P. 823



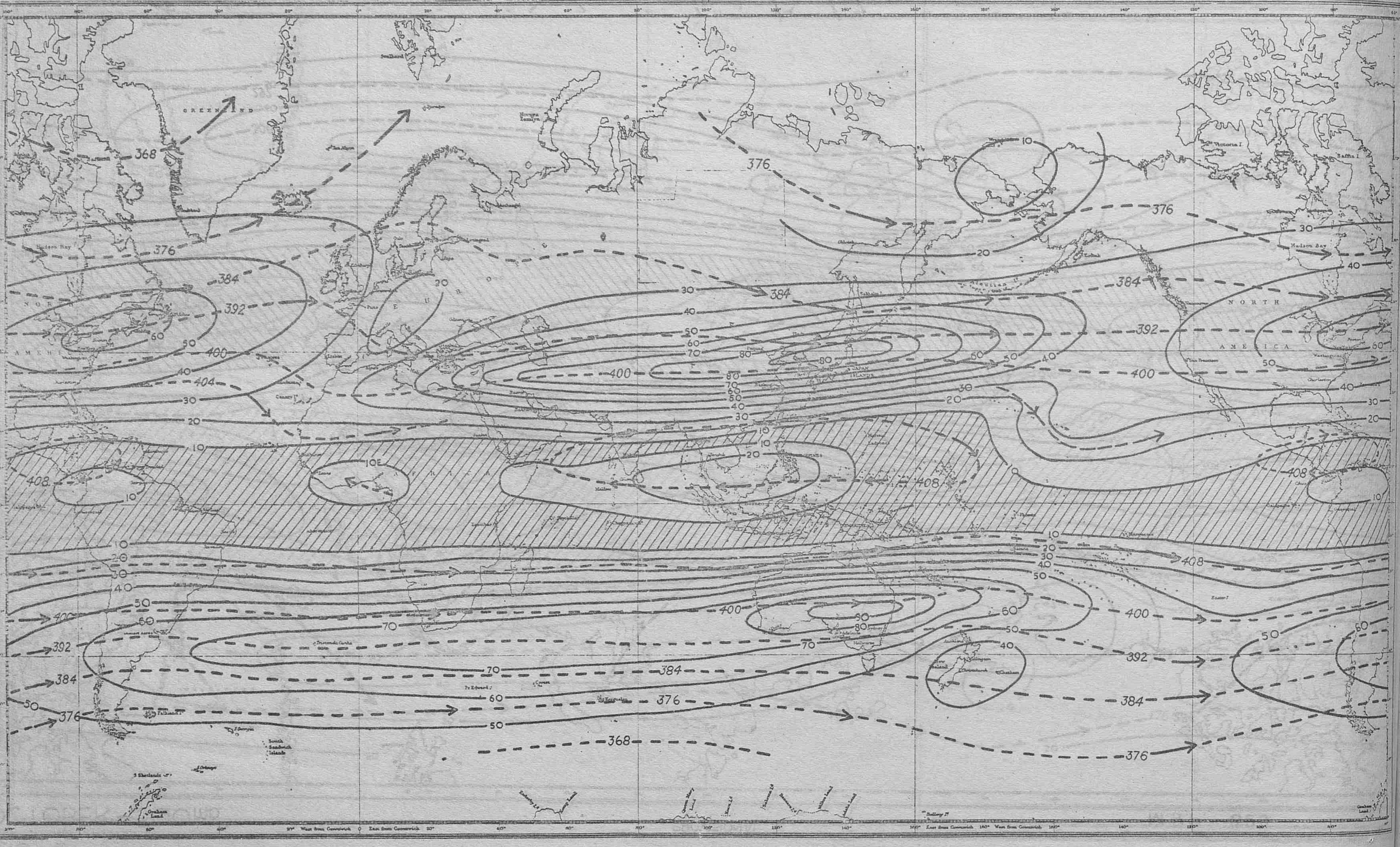
284
GREENLAND
Hudson Bay
292
300
308
316
318
284
10
20
30
40
50
60
70
80
10W.
10E
Borneo
JAPAN ISLANDS
SOUTH

FIG. 14

OCTOBER — 200 mb.

PROVISIONAL

M.R.P. 823



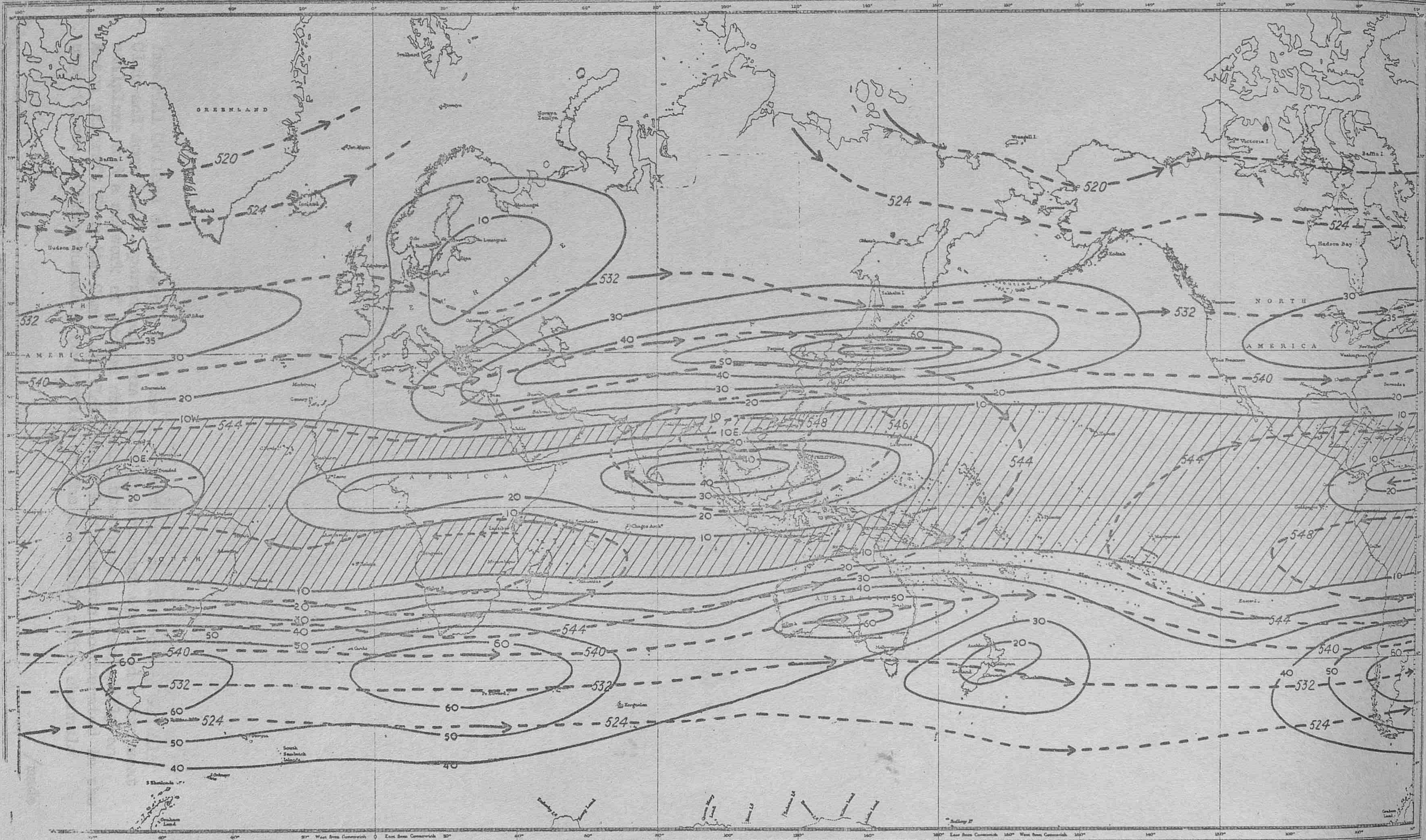
GREENLAND
EURO
NORTH AMERICA
AMERICA
JAPAN ISLANDS
Hudson Bay
368
376
384
392
400
404
408
10
20
30
40
50
60
70
80

FIG. 15

OCTOBER - 100 mb.

PROVISIONAL

M.R.P. 823


GREENLAND
EUROPE
AFRICA
NORTH AMERICA
SOUTH AMERICA
AUSTRALIA
Hudson Bay
Baffin I.
Iceland
Svalbard
Novaya Zemlya
Victoria I.
Wrangell I.
Sakhalin I.
Kerguelen
South Sandwich Islands
S Shetlands
Graham Land
520
524
532
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546
548
10
20
30
35
40
50
60
10W
10E

FIG. 16

[illegible]

A copy is available in the Library of the Meteorological Office."

AD35002

# AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

## THE CUMULUS CLOUD INVESTIGATIONS MADE BY THE METEOROLOGICAL RESEARCH FLIGHT DURING THE PERIOD AUGUST 4TH - 15TH, 1953

by

[illegible]

C.J. Aanensen (Meteorological Office)

### Summary

The results of eight flights made to investigate the structure and environment of selected cumuliform clouds are presented. Details are given of the aircraft's instrumentation for this work and of the techniques used.

The flights reported were arranged, as far as possible, to coincide with other cumuliform cloud investigations made by the Department of Meteorology, Imperial College.

1. **Introduction.** At the request of the Department of Meteorology, Imperial College, the Meteorological Research Flight participated in an investigation of cumulus type clouds during the period August 4th - 15th, 1953.

It was agreed that as far as possible, during this period, the Meteorological Research Flight would make at least one flight a day in the region of Cranfield, the location of Imperial College's experiments which were termed "Project Cumulus".

On most days a detailed sounding of temperature and humidity up to cloud tops was made in a Hastings aircraft and was continued with an investigation of the structure of a selected cloud and its immediate environment. As far as possible measurements were made within the cloud, at several levels, of temperature, water content, icing rate, vertical motions and turbulence, and samples were taken of water droplets and ice crystals. Observations were then continued below cloud base with traverses at regular height intervals to near the surface in which measurements of temperature, humidity fluctuations and turbulence were made. Photographs were taken of interesting cloud features.

Concurrently with the Meteorological Research Flight measurements, the

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London),

M.R.P. No. 826 by Mr. R.J. Murgatroyd and G.J. Day

A copy is available in the Library of the Meteorological Office."

M.R.P. 826
S.C.III/158
21st July, 1953.

AD 35002

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# AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

## The cumulus cloud investigations made by the Meteorological Research Flight during the period August 4th - 15th, 1952.

by
G.J. Day
R.J. Murgatroyd (Meteorological Office)

### Summary

The results of eight flights made to investigate the structure and environment of selected cumuliform clouds are presented. Details are given of the aircraft's instrumentation for this work and of the techniques used.

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Concurrently with the Meteorological Research Flight measurements, the Imperial College contingent, with the co-operation of the Surrey Gliding Club and some R.A.F. personnel from R.A.E. Farnborough operated a two seat sailplane and a network of ground observing stations. Occasionally an Anson aircraft from the Cambridge University Air Squadron co-operated in the experiments. Unfortunately congestion of the available radio channels made it impossible for the measurements made by the Meteorological Research Flight to be closely co-ordinated with those made by the other participants and the investigations while simultaneous were in fact made entirely separately. One of the difficulties in this work was the necessity of avoiding Air Ways and maintaining quadrantal heights during the cloud flying. Throughout all these flights continuous contact was maintained with the Meteorological Radar Station at East Hill.

These eight flights were also made as part of a long series of cumulus and other cloud investigation flights by the Meteorological Research Flight and this is therefore an interim report which has been produced at this stage to make these observations available at the same time as those of the Imperial College team. The first objective of the longer series of flights is also to produce and present measurements made in this country of the main cloud parameters as this is necessary before any checks can be made on the large amount of theoretical work that has been done on this subject.

## 2. Aircraft Instrumentation.

(a) Temperature measurements were made using the standard Meteorological Office flat plate resistance thermometer with a balanced bridge.

The readings made with this instrument must be corrected for kinetic or frictional heating and, when in cloud, for wetting of the element. The former correction may be made accurately in clear air but the magnitude of the correction when the element is wet is somewhat doubtful. Recent American work indicated that the dry air correction should be reduced in the ratio saturated adiabatic lapse rate to dry adiabatic lapse rate and this is the procedure that has been adopted. Little confidence may be placed in the temperatures recorded immediately after leaving cloud since there is an undetermined wet bulb effect operative until the bulb of the thermometer has dried, a process normally taking about 30 seconds.

The lag of the thermometer is of the order of 8 seconds and the accuracy of the temperature readings about ±0·2°F in clear air. In and after leaving cloud there may be errors of about 1°F.

(b) Humidity measurements were made using a manually operated Dobson-Brewer Frost Point Hygrometer.

There are no corrections to be made but the instrument is slow to operate above frost points of about 0°F. A skilled observer may make readings once or twice a minute to an accuracy of about 1°F but below -50°F the speed of operating drops to about once every two minutes and the accuracy 2-3°F.

(c) Icing rate was measured using a rotating disc icing meter. A disc about 4 cm. in diameter and 2 mm. thick rotating at 2 r.p.m. is exposed to the airstream. Bearing on the edge is a probe which is mechanically connected to a transducer. Deflection of the probe by ice accumulated on the edge varies an air gap and thus an inductance in a previously balanced a.c. bridge circuit. Out of balance current is rectified and displayed on a meter as amount of ice per unit volume swept out by the disc.

The instrument is calibrated by exposure to standard conditions in an icing wind tunnel and its readings then converted to cloud water content. Small errors arise out of the departures of the density of the ice collected from that obtaining in the tunnel and out of inaccuracies in the standardisation of the tunnel. In addition the readings are only comparable when the cloud is purely supercooled water droplets. However, until recently, the cloud water contents measured with this instrument were always considered to be within say ±20% accuracy. Recent work, however, by Ludlam [1] in this country and by members of the N.R.C. Canada [2] has demonstrated that the instrument fails completely to give correct water contents when the actual cloud water content is greater than about 2 gm/cu. metre within the usual aircraft icing temperature range.

(d) Water content was also measured by means of a heated cylinder water content meter. It comprises two cylinders about a centimetre long and 0.3 cm. in diameter. One cylinder is exposed to the cloud droplets and the other is shielded from them in such a manner that both are under the same conditions of air flow.

The cylinders are wound with temperature sensing elements which form two arms of a Wheatstone bridge network. This is balanced in clear air in order to compensate for minor differences between the cylinders. Out of balance current, when in cloud, resulting from expenditure of power in the evaporation of intercepted droplets is recorded as water content of unit volume of air swept out by the cylinders.

Errors of the instrument arise mainly out of the calibration in the icing tunnel which is performed at the same time as that of the rotating disc icing meter. Accuracy is lowest at low water contents. In general the error is probably about ±10% - 15%. Lag is about 10 seconds.

(e) Turbulence was estimated subjectively every 10 seconds on a scale of 0.1 - 1.0g which has previously proved quite accurate for experienced observers. An accelerometer was also available but rarely used as very little extra information on bumpiness would have been provided by its use.

(f) Water droplets were sampled by both the Mazur type impactor and a recently developed pneumatically operated multi-shot impactor. In the former a magnesium oxide coated microscope slide, carried in a carefully designed housing, is exposed to the airflow for about a hundredth of a second by a simple shutter system. Intercepted droplets make craters in the coating whose size bear a known relation to the size of the droplets making them. The slide is viewed under a microscope and counts are made. Errors arise in the derived size distributions when the integrated area of the craters exceeds a certain fraction of the exposed area. An additional but small error is due to the low collection efficiency of the impactor for drops of diameter less than about 5.0 μ.

The pneumatic impactor has five slides which are carried on pistons. Operation of a valve pushes the slide out into the airflow after which it returns under the action of a spring; the time of exposure depends on the operation of the valve and at present is not accurately timed. The slides are coated with a heavy oil-Spirex 230-possessing desirable viscosity characteristics and in which the intercepted droplets lie unmodified. After exposure the slides are microphotographed in the aircraft. Errors are similar to those inherent in the Mazur method but efficiency of catch for smaller droplets is higher.

(g) Ice crystals are sampled by the Schaefer replica method. A microscope slide cooled to a sub-freezing temperature and coated with a 5% solution of Formvar 15-95 in ethylene dichloride is exposed to the decelerated airstream. Ice crystals falling on the slide form a transparent plastic skin round themselves and a hollow replica of the crystal is left attached to the slide. The replicas are viewed under a microscope by dark ground illumination. The only probable errors which may occur are due to shattering of the crystals on collection but care has been taken to prevent this occurrence as far as possible. Typical examples are shown in Figs. 20 a-i.

/(h)

(h) Vertical motions in cloud are derived from a record of the aircraft's height and airspeed. For all traverses the pilot was instructed to fly the aircraft as far as possible with constant power and elevator settings. Should the speed and height then increase together the aircraft must be in an updraught whose minimum strength recorded over the time lapse since the last reading is given by the displacement. Similarly mean minimum downdraught strength may be obtained. Accurate assessments of mean draught strengths may only be obtained when there is a height displacement at constant airspeed. Any change in height in antiphase with a change in airspeed is interpreted as caused by a control movement on the part of the pilot.

(i) The temperature field below and around cumulus clouds can be examined in great detail by means of the Ultra Rapid Thermometer described in M.R.P. 658[3]

Rapid variations in the wind field are similarly examined by means of hot wire anemometer apparatus using similar circuit arrangements. By means of a single hot wire rapid variations in total wind are recorded and, by means of a differential head, vertical components can be studied.

Very few records were obtained with this apparatus during these particular flights due to various instrumental faults but separate reports are in preparation both on the instruments themselves and recent studies made with them on the detailed wind and temperature field in which cumulus clouds are found.

3. Operational Procedure. After take-off the aircraft normally climbed to about 5,000 feet on track for Cranfield. When at this height instrument tests and calibrations were made.

The pilot then established communication with both East Hill Radar Station and Cranfield Control and the aircraft was identified and accepted for surveillance by the former.

At Cranfield the aircraft descended to 1,000 feet ICAN and commenced a temperature and humidity sounding, observations being made normally at 500 feet vertical intervals and continued well above the tops of any convective cloud in the area.

During the climb the pilot obtained information from Cranfield Control about the conduct of the ground experiment and determined the area in which the ground organisation was most interested. The radio channels however were invariably too congested to allow any close co-operation.

At the conclusion of the sounding the aircraft took up position over a selected cloud usually a large growing cumulus and made traverses in clear air above it and then through it, usually at five heights intermediate between tops and bases.

On reaching cloud base observations with the hot wire anemometers and ultra rapid thermometer were commenced. Usually at least three clear air traverses were made with each instrument serviceable, one being used at each height in succession before selecting the next.

At the completion of the investigation the aircraft returned to base, minor investigations being undertaken whilst on track.

4. Treatment of Results. In paragraphs below each flight is described in detail under three main headings:-

(i) The execution of the flight.

(ii) The soundings and their relation to the cloud features observed.

(iii) The results of the traverses.

Photographs of the clouds and also microphotographs of the cloud droplets and crystals are given in some cases.

It is not possible to describe all the cross section runs in detail within the compass of a short report so sub-para. (iii) has been usually divided into five sections (a) cloud structure (b) composition (c) vertical currents and turbulence (d) icing and water contents and (e) field below the cloud. In each case only a short description has been given. Full details have been tabulated but owing to their volume have not been attached to this report. In the more interesting cases, however, abridged tables of the main cloud features and tables and diagrams of the droplet and crystal distribution have been included.

## Summary of Observations

Flights 1, 2, 3, and 8 were made in large cumulus or cumulonimbus clouds and the remainder in small cumulus or stratocumulus clouds.

When there were cumulonimbus clouds in the area the soundings were made up to about 22,000 feet usually although this was not as high as some of the cloud tops. Above this height the rate of climb of the Hastings aircraft becomes very slow. On the other occasions the soundings were made to 10,000 feet which was well above the cloud tops.

### (i) The soundings and their relation to observed cloud features

(a) The height of the cloud base was generally in good agreement with that expected from the "parcel" method used in conjunction with the sounding. The height of cloud tops could not be estimated with the same accuracy by this method, although the main features were apparent e.g. it was quite obvious from each sounding whether it was a cumulonimbus or a small cumulus occasion. However, from a practical point of view the difficulty of forecasting the various cumulus parameters appears to be mainly a question of forecasting the sounding rather than a question of whether the parcel, slice, or similar methods are adequate to determine the amount of cumulus development. Although they are not reproduced here, the nearest radio sonde ascents were compared with the aircraft ascents and large differences were noted at times. These were, no doubt, due mainly to differences in space and time of the two soundings combined with developments of the synoptic situation. It appears from the aircraft soundings and cloud observations that a variation of 1°F in the ambient temperature at any level is enough to determine whether the cloud top shall be at that level or several thousand feet higher and the distribution of the tops of the smallest clouds seemed to be determined by "stops" of this magnitude. Similar differences will also determine whether there will be layers of stratocumulus or altocumulus mixed in with the cumuliform cloud at given levels. Hence a forecast of the sounding to this accuracy appears to be necessary before the detailed cloud forecast can be given.

(b) Although it appears inevitable from the water content measurements that entrainment must be an important factor in cumulus development several cases were noted of large cumulus clouds ascending through a comparatively dry environment. Any large scale mixing should then have destroyed the cloud tops. Our present view of entrainment from these and other flights is that it is a factor of considerable importance near the base and perhaps for the lower half of a large cumulus cloud but that at higher levels its magnitude decreases considerably.

(c) The tops of stratocumulus layers were usually determined by a decrease in the lapse rate or inversion, combined with a decrease in humidity. The bases were often not related to any obvious feature in the sounding although wind shear may have been the most important factor in these cases.

/(ii)

### (ii) Cloud structure

(a) The large cumulus or cumulonimbus clouds investigated had lateral dimensions of about 10 miles at their middle levels. Lower down they often became larger with small cumulus clouds alongside them and at greater heights decreased in extent. Thin stratocumulus layers sometimes of lenticular form were often observed at various levels between the cumulus clouds. The large cumulus clouds all consisted of several cells in different stages of maturity and the individual cells varied in extent from one to four miles. The state of maturity of the different cells was often very obvious e.g. in the tops of a cumulonimbus one cell was glaciated while another contained considerable supercooled water. On other occasions some cells produced precipitation whilst adjacent cells did not do so and the direction of the vertical currents changed between and across the cells. Cloudy but relatively inactive regions up to a mile across were sometimes noted between the cells. The younger cells were usually downwind of the more mature cells.

(b) The small cumulus clouds usually had only one cell and were one to two miles across.

### (iii) Cloud composition

(a) In the layer clouds ice crystal samples were taken at temperatures between -9°F and freezing point. On Flight No. 3 samples were also taken in altostratus cloud at about 20,000 feet. In Flight No. 8 no ice crystals were found down to temperatures as low as +6° F at 17,000 feet although it appeared later that glaciation of the cloud tops started soon after this observation was made.

The nature of the crystal distributions was very confused and details are given in the Tables relating to the individual flights. Dendritic types were rare and it appears that this type must be confined to a very narrow temperature band. Single hexagonal and tetragonal plates were most common with a few columns. Both the individual crystals and the aggregates appeared to decrease in size towards the top of the clouds.

Both snow (between 22°F and 28°F) and hail (at 22°F) were reported in Flight No. 3 only. Precipitation static was observed between temperatures of + 12°F and +23°F in a mixed cloud in Flight No. 1 and also reported at +42°F in rain in Flight No. 2.

(b) Droplet samples were obtained on every flight except the first and showed considerable variation both horizontally and vertically throughout the clouds. Tables and diagrams of the distributions are given in connection with the descriptions of the individual flights. Usually droplets about 100 μ and 200 μ diameter can be clearly seen on the aircraft's windscreen and rain is then reported. Hence in the case of the large clouds where ice crystals were present and the Bergeron process possible drop sizes of 100 μ diameter or more were expected and in fact were seen on the windscreen and were also measured by the sampling process. But in Fights 6, 7 and 8 which were through a cumulus cloud extending above the freezing level but not containing ice crystals droplets of this size were also measured. In addition large droplets were found during Flight 4 in a cumulus cloud which did not reach the freezing level and in Flight 5 which was made through small cumulus cloud intermixed with stratocumulus sheet with tops well below the freezing level. Small droplets were also of course always measured so that the droplet spectrum in all cases was very wide. This indicates that the coalescence process was generally very active in all the cumulus and stratocumulus clouds examined. In all cases there were reports of rain or drizzle within and below the cloud although it may not always have reached the ground. Only in Flights 1, 2 and 3 would it have been possible for the Bergeron process to have been the primary cause of the rain drops. The air was predominantly maritime in origin throughout the period of these experiments so that sea salt nuclei would almost certainly have been available for the necessary condensation process to form large droplets which would initiate the coalescence process. The large horizontal variations of droplet spectra are ascribed to different stages of the coalescence process being reached in different cells in addition to the effect of sampling at different points in the same cell.

(iv) Vertical motions and turbulence

(a) Vertical motions averaged over distances of about half a mile were usually less than 5 feet per second even in the larger clouds. Occasionally, however, values of up to 20 feet per second were recorded.

(b) Turbulence was very variable. In most cases values of 0.3g (a value which makes writing or movement in the aircraft difficult and which is very uncomfortable if maintained for a period of time) were the maxima encountered in and below the clouds. Frequently the maximum value in the traverses was only 0.1g. Bumpiness was encountered in new hard towers at the tops of the clouds while the traverses through the crystal portions and anvils of the cumulonimbus clouds were generally quite smooth. The roughest conditions occurred in the precipitation areas about half way up the clouds both above and below the freezing level. Thus Flight No. 3 gave bumps of 0.5g in rain at 9,000 feet temperature 36°F in a cloud with base at 3,000 feet, tops 15,000 feet. Bumps of 0.7g were observed in Flight 8 at 13,000 feet temperature 23°F and 15,000 feet temperature 12°F in a cloud with base at 3,000 feet and tops 17,000 feet. It is interesting to note that in this latter flight only large water droplets were present and no change to ice crystals had yet taken place when the measurements were made. The largest bumps usually coincided with the probable cell boundaries or start of an up or downdraught. There was a tendency for the bumps to become more frequent towards the base of the cloud.

(v) Water content and icing

(a) Both water content and rate of icing tended to rise towards the centre of the cells. There was also a tendency for water content to rise with altitude except near the tops where it fell away rapidly. Water content always varied very rapidly with distance across a cloud, but it tended to be more uniform at the lower levels. Even its peak value very rarely approached that which would be calculated from adiabatic lifting considerations so that it appears certain that at some stage mixing or entrainment must be a very important factor in the development of the cloud.

(b) Icing was reported at temperatures between -9°F and 32°F. The former observation was taken in a hard cumulus top at 22,000 feet adjacent to a cell which had already developed an anvil in which there was no icing. In the same hard top at 19,000 feet and +2°F a maximum value of 2 gm/cu.metre was recorded. In another case 1.2 gm/cu.metre was found at 13,000 feet temperature 22°F during a traverse in which crystals, snow, hail and droplets were all observed.

(vi) The field below the cloud base

The work on this aspect was disappointing during these particular flights due to instrumental failures. More successful experiments have since been made and will be reported elsewhere. Some features however were apparent and these are listed below:-

(a) The vertical motions were usually smaller than in cloud and became more coherent from the ground upwards. The cell size was smaller than in cloud being usually half a mile to a mile in horizontal extent.

/(b)

(b) Temperature tended to rise in rising air and fall in descending air, particularly close to cloud. The temperature excess was about 0.3°F and the deficiency somewhat smaller.

(c) Below cloud well marked downdraughts were usually associated with rain areas.

## Discussion

The object of this type of work is to gain a knowledge of the various cloud parameters and how and why they develop during the life of the cloud.

A certain amount of progress has been made in the instrumental field towards that end as water content, icing rate, vertical motions and turbulence can now be measured whilst droplets and crystals can readily be sampled in flight. There are however several large gaps in the instrumental equipment for this work such as accurate temperature measuring devices in cloud, instruments for measuring rain drop size and water content, instruments for measuring water content in ice crystal cloud etc. and work is continuing to develop these and also to improve existing instruments.

It appears that the present technique of measurement is inadequate as the life cycle of a given cumulus cell is usually about half an hour and the method of five or six successive traverses at different levels by the same aircraft takes about an hour. This means that the observations at one level are taken at different stages of a given cell's development from those at other levels. A fuller investigation of a cloud would include simultaneous measurements at several levels at about 5 minute intervals over a period of an hour or so. This of course would involve the use of several aircraft. At present this is not possible but a new series of flights has now been started in which these successive measurements at a given level are made with one aircraft. This will give information on variations with time at given levels and supplement the present series.

As the eight flights described in this report are part of the larger series it is premature to draw any firm conclusions from them. Several points however are of interest and are listed below:-

1. The large effect of small changes in environment temperature on the cloud development.

2. The necessity for entrainment or mixing to explain the water content of clouds but the likelihood that it is only important at their lower levels.

3. The presence of large droplets and wide droplet spectra in clouds where the Bergeron process could not have been in operation and the consequent importance of coalescence as a factor in the development of rain.

4. The low frequency of occurrence of dendritic crystals compared with prismatic and columnar types.

5. The data on structure and size of the clouds, icing, water contents, vertical currents and turbulence is consistent with that obtained by other investigators.

## Acknowledgments

Acknowledgments are due to F.O. Bavin, R.A.F., who flew the Hastings aircraft used during these flights. Though inexperienced on the type he flew with accuracy under what were often difficult circumstances.

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/Flight 1.

## Flight 1. 5/8/52.

1. Execution of the Flight. An ascent was made to 22,000 ft. during which temperature and frost point was measured every 1,000 ft. from 1,000 ft. ICAN. Various cumulonimbus tops extended to an estimated height of 30,000 ft. or more and penetrated a sheet of cirrostratus, particularly in the east.

It was decided to investigate a maturing cumulonimbus whose top was about 25,000 ft. and three minute traverses were made through its centre at 22,000, 19,000, 16,000, 13,000 and 10,000 ft. It was found that dissipation had occurred at the base which by this time - about 30 minutes after the start of the experiment - had risen to about 9,000 ft.

The aircraft then returned to base.

2. The sounding and its relation to observed cloud features.

(a) Results of the aircraft ascent and the winds from the afternoon Larkhill radio sonde ascent are plotted on Fig. 1. Estimated surface temperatures and humidities for the Cranfield area at the time are also given. This ascent lay roughly along the 60°F saturated adiabatic line and was quite moist.

(b) Considering Fig. 1 it appears that:-

(i) The lowest cloud base was 960 mb. which by the parcel method would be produced by a surface temperature of 70°F and dewpoint 62°F. The actual surface temperatures over the area were 67°F and dewpoints 64°F at the time.

(ii) With a condensation level of 960 mb. and the temperature ascent having a lapse rate rather more than the saturated adiabatic lapse rate, cumulonimbus clouds to very great heights might be expected. A photograph of the cloud actually investigated is given in Fig. 2b.

(iii) These cumulonimbus clouds were in fact encountered and as there are considerable dry layers around 750 mb. and above 600 mb. there could not have been any considerable entrainment at these levels.

(iv) The beginning of the lower dry layer coincided with the top of a stratocumulus layer. Thin stratocumulus cloud layers mixed in with the cumulus clouds might have been expected at about 800 mb. and 600 mb. from considerations of stability and humidity. In fact the only well marked stratocumulus layer occurred at about 800 mb. where well marked lenticular shapes were observed (Fig.2a).

The stabilising of the lapse rate from 910 mb. to 875 mb. coincides with the tops of small cumulus clouds.

3. Results of the traverses. The main cloud features are summarised in Table 1 and short notes on these features follow:-

(a) Cloud structure. The cloud under investigation had lateral dimension 10 miles and contained two cells at most levels all of lateral dimensions 2 - 3 miles. Of these the older cell lay to the south, i.e. upwind, and was almost entirely composed of ice crystals; from it was generated the anvil. The most active, northern, cell contained supercooled water droplets and caused aircraft icing.

(b) Composition. Ice crystals were sampled at all heights and a detailed description of the results is given in Table 2. The variation in type, after allowance for the presence of vertical motions, roughly follows that suggested by Nakaya[4]. The type was predominantly single plates with a few columns and there were indications of an increase of the size of single plates and aggregates towards the cloud base.

Measurements of droplet size were unsuccessful, probably due to the destruction of the MgO slide coating by ice crystals in the upper-regions of the cloud; in the lower regions no attempt was made due to the risk of lightning strikes. This risk was considerable as precipitation static was strongly evident on the aircraft V.H.F. system. No direct observations were made of precipitation but precipitation static was encountered at levels below 16,000 ft. in both the old cells.

(c) Vertical motions and turbulence. No violent up or downdraughts were measured, the normal value, averaged over half mile intervals, being 5-8 ft./sec. Turbulence was moderate and never exceeded an estimated 0.3g. It was usually experienced at the boundary between vertical currents of opposing sense and was most noticeable in the lower regions of the cloud.

(d) Water content and icing. Icing was encountered only in the northern cell and at temperatures between -11 and +12°F. Water content varied considerably both horizontally and vertically throughout the cloud. Values of 0.7 g/m$^3$ were measured at the base and a peak value of 2.0 g/m$^3$ was recorded by the icing meter at 22,000 ft.

(e) Field below cloud base. No observations were made below cloud base.

Table 1.

Main cloud features of decaying cumulonimbus investigated in Flight 1.

Cloud Traverses

| Height feet | Mean Temp. °F. | Turbulence Max g | Turbulence Mean g | Icing Max g/cu.m. | Icing Mean g/cu.m. | Water Content Max g/cu.m. | Water Content Mean g/cu.m. | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|---|
| 22,000 | -9 | .3 | .1 | .85 | .1 | - | - | 7 | 2 | Cell 1 Droplets<br>2 Crystals |
| 19,000 | +2 | .1 | .1 | 2.1 | .6 | - | - | 9 | 2 | Cell 1 Droplets<br>2 Crystals |
| 16,000 | +12 | .2 | .2 | un-serviceable | | .6 | .4 | 10 | 1 | Mixed Droplets and Crystals Precipitation Static. |
| 13,000 | +23 | .3 | .2 | un serviceable | | .2 | .1 | 4 | 2 | Cell 1 Droplets<br>2 Crystals<br>Precipitation Static |
| 10,000 | +32 | .3 | .2 | - | - | .7 | .3 | 5 | 1 | Mixed Droplets and Crystals. |

/Table 2.

Table 2.

Analysis of cloud elements - Flight 1

| Height ft. | Temp. °F. | Single Crystals |  |  |  |  |  |  |  |  |
|---|---|---|---|---|---|---|---|---|---|---|
|  |  | Plates |  |  | Columns |  |  | Aggregates |  |  |
|  |  | Type | Size | Remarks | Type | Size | Remarks | Type | Size | Remarks |
| 22,000 | -11 | Hexagonal and tetragonal | 50-210μ | 1:100 hollow | Hexagonal oval | 420 x 70μ |  | Branched chains | 350-700μ |  |
| 19,000 | +1 | Hexagonal and tetragonal, half each | 50-270μ | Hexagons rather bigger |  |  |  | Chains of drops and crystals 1:100 have stellar tendencies | 350μ | 1:20 of smaller tetragons had 80μ droplets on their points. Drops often branch without merging. |
| 16,000 | +12 | Hexagonal and tetragonal in ratio 20:1. | 20-350μ | Tetragons have preferred direction of growth. |  |  |  | Chains, occasionally closed. | 1100μ | 1 in 100 tetragons had 80μ droplet on point. 1 in 150 dendritic. |
| 13,000 | +22 | Hexagonal and tetragonal ratio 1:5. | 20-170μ |  |  |  |  | Irregular, a few unbranched chains. | 900μ mainly but some 500-1000μ |  |
| 10,000 | +31 | Hexagonal and tetragonal | 35-100μ | Some had stellar hollows in centre. | Capped columns | 350 x 100μ, caps 140μ | Occasionally 200 x 120, caps 170 | Mainly massy aggregates. | 700-1000μ | Some dendrites, some individual hexagons of 350μ with preferred direction of growth. |

Flight 2. 6/8/52.

1. Execution of the Flight. During the morning a Mosquito aircraft made a temperature sounding above Cranfield, and also observed the heights of the bases and tops of the various cloud layers.

In the early afternoon the Hastings aircraft made a further sounding over Cranfield to 18,000 ft. ICAN. Thereafter a mature cumulus congestus was selected for investigation and traverses made through it at 18,000 and 15,000 ft. and the cloud then abandoned because of rapidly advancing decay which appeared to be occurring from the base upwards. A new cloud was selected which was growing in the Kettering area and traverses made above at 14,000 ft. and then through it at 13,000 ft. and subsequently at 2,000 ft. vertical intervals to 5,000 ft. Below cloud base runs were made with the ultra-rapid thermometer and the non-directional hot wire anemometer at 3,000, 2,000 and 1,000 ft. respectively.

2. The ascents and their relation to cloud features. The aircraft ascents and the winds from the afternoon Larkhill radio sonde ascent together with the estimated surface temperature and dewpoints in the Midlands are given in Fig. 3 and Fig. 3a.

Considering Fig. 3 it appears that:-

(a) The lowest cloud base was 915 mb. in the afternoon which is consistent with an estimation based on the parcel method and a surface temperature of about 74°F and dewpoint of 62°F. The actual values in the area that afternoon were temperatures 68°F to 70°F and dewpoints about 63°F.

During the morning the Mosquito observations showed small cumulus from about 950 to about 920 mb., a thin stratus layer at 880 to 850 mb., a stratocumulus layer from about 810-795 mb., some altostratus at about 610 mb., with some cumulus tops growing through it, and in the distance, both to east and west, large cumulus and cumulonimbus clouds were seen. In addition there was 7/8 of cirrus or cirrostratus cloud above 450 mb. so there can have been no lack of ice crystals in the atmosphere at high levels.

(b) Cumulus forming at the condensation level of 915 mb. in the afternoon could readily ascend to heights above the maximum of the aircraft sounding and this was confirmed by observations of many large cumulus and cumulonimbus clouds as well as being expected from the Mosquito observations.

Some photographs of these clouds are shown in Figs. 4a, b, c, and d.

(c) The humidity sounding shows a comparatively dry layer at about 700 mb. If there was any entrainment at this level it apparently made very little difference to the development of the cumulonimbus clouds.

(d) One stratocumulus layer was observed at about 760 mb. just below the comparatively dry layer. This layer can be seen clearly in Fig. 4c. The reason for its existence is not obvious from the tephigram although humidity was generally high at this level, and the wind is generally changing there from a southeasterly to a southwesterly regime.

3. Results of the traverses made. The main cloud features are summarised in Table 3 and short notes on these features follow:-

(a) Cloud structure. The cloud investigated appeared to consist of three cells, 2 miles, 4 miles and 2 miles wide respectively. The southern (upwind) cell was in an advanced stage of growth and contained a downdraught and rain. Dissipation appeared to be taking place from the base upwards but this may have been a result of the "direction" of investigation - from the top downwards. The central cell was mature and contained both up and downdraughts, with rain in the updraught, whilst the northern cell was small and probably in an early stage of growth. There are indications that the 7,000 ft. traverse was rather off course and did not pass through this cell.

(b) Composition. Ice crystals were found only at 11,000 ft., apparently in the central cell, and they were mainly hexagonal columns and tetragonal plates. This was a region of mixed crystals and droplets. As cloud top temperature was 23°F presumably some natural seeding was taking place from other ice clouds.

Droplets were sampled at all heights but there were no systematic trends in sizes, presumably due to the random manner in which sampling was done relative to the cells. Some very large droplets of diameters up to 500 $\mu$ were observed in the 7,000 - 9,000 ft. region. Details of the cloud elements are given in Table 4.

(c) Vertical currents and turbulence. Up and downdraughts averaged over about half a mile were usually less than 10 ft./sec. but an updraught decreasing in strength upwards from 10-15 ft./sec. was observed in the rain area of the central cell. Downdraughts of about 5 ft./sec. were encountered in the run over the cloud particularly on the north side. Turbulence was quite moderate at all levels being less than an estimated 0.3g; in all but the lowest levels it could be associated with the shear zone between vertical motions of opposing sense.

(d) Icing and water content. No icing was recorded or reported though it seems likely that there may have been slight icing in the top of the mature cell, where the rate of icing may have been below the capacity of the rotating disc meter.

Peak water content was 0.8 g/m$^3$ and at 9,000 ft. Water contents in the lower regions were more constant than in the higher.

(e) Field below cloud.

3,000 ft. At this level a rain area 2 miles wide apparently associated with a single cell was encountered. Vertical motions were generally less than 5 ft./sec. Downdraughts were strongest in the rain area where they had apparent temperature deficiencies up to 2°F but the most common excess or deficiency in the draughts was only about 0.1°F. Temperatures tended to be higher under firm cloud bases and this was also the most common updraught region. Gusts were usually very small except for one sharp edged gust at the edge of a downdraught associated with the rain area.

2,000 ft. This was a region of small and apparently random vertical motions less than 5 ft./sec. with temperature variations usually less than 0.2°F. There was however one variation of 0.5°F associated with an updraught. Downdraughts and lower temperatures tended to coincide with breaks in the cloud above, and were again most evident in the rain area. A sharp edged gust was recorded at the edge of a 5 ft./sec. downdraught at the edge of an overcast region.

1,000 ft. Vertical motions were more pronounced at this level and reached about 10 ft./sec. Temperature pulses of ±0.2°F were frequent and usually had a horizontal extent of about 300 yards. On one occasion there was a downdraught of 5 ft./sec. under cloud with an updraught of 10 ft./sec. under the surrounding blue sky and this updraught appeared to have a temperature excess of 0.4°F. Gusts of 5 to 10 ft./sec. were general and were most frequently found at the edges of up or downdraughts.

Table 3.

## Main cloud features of maturing cumulus congestus investigated in Flight 2.

Cloud Traverses

| Height feet | Mean Temp. °F. | Turbulence | | Icing | | Water Content | | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|---|
| | | Max g | Mean g | Max g/cu.m. | Mean | Max g/cu.m. | Mean | | | |
| 13,000 | 23 | .2 | .2 | - | - | .2 | .2 | 1 | 1 | Droplets. |
| 11,000 | 30 | .2 | 0 | - | - | .5 | .1 | 2 | 1 | Mixed droplets crystals and rain. |
| 9,000 | 36 | .3 | .2 | - | - | .8 | .6 | 6 | 3 | Droplets and rain. |
| 7,000 | 42 | .3 | .1 | - | - | .6 | .2 | 4 | 2 | Droplets and rain. Precipitation static. |
| 5,000 | 49 | .2 | .2 | - | - | .7 | .3 | 5 | 2 | Droplets and rain. |

Table 4.

## PROJECT CU. - FLIGHT 2

### Maturing cumulus congestus

| Height feet | Temp. °F. | Water Droplets | Ice Crystals |
|---|---|---|---|
| 13,000 | +24 | 14-40 μ, mainly 20μ. Density about 300/cm$^3$. | Nil. |
| 11,000 | +29 | 7-30μ, some 55-110 μ. Density about 300/cm$^3$. | Hexagonal columns 50 x 30μ, and tetragonal plates. |
| 9,000 | +36 | 7-30μ and 55-110 μ. Density about 300/cm$^3$. | Nil |
| 7,000 | +42 | 7-55μ, but some 80-500 μ, of which most are $<$ 350μ. Density about 200/cm$^3$. | Nil. |
| 5,000 | +49 | Mainly 7-10μ. Density about 300/cm$^3$. | |

/Flight 3.

Flight 3. 7/8/52.

1. Execution of the Flight. During the afternoon a sounding to 22,500 ft. ICAN was made over Cranfield in the Hastings aircraft and a developing cumulus lying to the north of Cranfield subsequently chosen for investigation. Traverses were made through it at 15,000 ft. and thereafter at 2,000 ft. vertical intervals to 3,000 ft. In addition crystal samples were obtained in an altostratus layer from 19,000 to 23,000 ft. at 1,000 ft. intervals.

2. The soundings and their relation to observed cloud features. The results of the soundings have been plotted on Fig. 5. Considering this diagram:-

(a) From parcel method calculations the lowest condensation level would be at about 930 mb. with a surface temperature of about 70°F and a dewpoint of 60°F. Observed surface values in the area were about 66°F and 59°F respectively.

The main cumulus cloud base appeared to be at about 930 mb. but fractocumulus was reported at much lower levels. At these very low levels it was very moist and condensation was observed over the aircraft wings.

(b) Upper air conditions on this ascent were rather more stable than on Flights 1 and 2. There are several small decreases in the lapse rate e.g. at 880 mb. (which agreed with the tops of the fractocumulus layer), 780 mb., 640 mb. and 510 mb. which would tend to stop vertical motion and also an altostratus layer above which must have reduced surface heating. This ascent was, however, made rather near to a front which was giving rain over England south of the Thames and it appears probable that the large cumulus clouds were due at least in part to general convergence in the area.

(c) There was a very dry layer at about 680 mb. while cumulus tops were extending to at least 600 mb. so it appears that, in this case also, entrainment at these levels is not considerable. The very dry layer started at the top of a thin altostratus layer.

(d) The nearness of the frontal cloud and the presence of the high altostratus layer must certainly have provided a good source of ice crystals for natural seeding in the area.

(e) Conditions were poor for aerial photography during this flight and the photographs obtained were not worth reproducing.

3. Results of the traverses. The main cloud features are summarised in Table 5 and short notes on these features follow:-

(a) Cloud structure. The cloud proved to be a complex of two cells, of which the most easterly -cell 1- was dissipating and the other -cell 2- reaching maturity. Cell 1 was surmounted by pileus when first investigated and in its upper portions regions of rain, snow and hail could be distinguished. Rain was confined to the most easterly portion and did not appear to reach the ground. The cell contained an updraught in the upper portion but a downdraught was observed later in the lower levels. Cell 2, which was separated from Cell 1 by a gap about half a mile wide, set in the shear zone between an updraught and a downdraught, was in the rain stage save for a small snow region in the uppermost portion of the cloud. The general rain intensity appeared to increase fairly regularly to a maximum near the centre of the cloud at all levels. The maximum lay in a fairly strong updraught and from this region rain appeared to reach the ground.

(b) Composition. Details of the droplet and ice crystal sampling in both the altostratus and the cumulus cloud are given in Tables 6 and 7.

The ice crystals, which were sampled in Cell 1 of the cumulus cloud in its upper 2,000 ft., i.e. at temperatures less than about +22°F, were predominantly hexagonal plates of diagonal dimension 20-170 μ in aggregates of about 700 μ diameter. Some dendritic structures were present, however, in the region of rain development in the extreme top of the cloud. Sometimes the aggregates contained droplets of diameter 10-60 μ attached to the ice crystals.

Water droplets were of approximately equal density - 700/cm$^3$ - at all heights and basically of fairly constant width of distribution. The majority of droplets sampled appeared to have diameters of less than 10 μ although 10-20 μ was also common. Large droplets of 100 μ diameter or larger were present at all levels. There was a maximum width of the superimposed rain spectrum at 9,000 ft.

The altostratus was composed of plate and columnar crystals. The latter were about 200 μ long and were more common at the higher levels. The former increased in size from 170 μ to 350 μ diagonal length between 23,000 and 20,000 ft. Some 14 μ diameter droplets were observed in the altostratus at 20,000 ft. where the temperature was -3°F.

(c) Vertical motions and turbulence. Vertical motions averaged over half a mile, did not normally exceed 10 ft./sec.

Turbulence was rather higher than usual, reaching a maximum estimated value of 0.5 g at 9,000 ft. and becoming more probable at the lower levels.

(d) Water content and icing. Icing was encountered only during the 13,000 ft. traverse and at about +23°F. A peak value of 1.2 g/m$^3$ was recorded.

Water content was quite variable at all levels, a peak value of 0.95 g/m$^3$ being recorded at 9,000 ft.

Visibility at the 9,000 and 7,000 ft. levels was often 5 yd. or less.

The hot cylinder water content meter although designed primarily to measure water content in cloud gave readings between 0.1 and 0.4 gm/m$^3$, in the rain area underneath the cloud. These must be regarded as minimum values as the collecting head of this instrument is too small to work efficiently as rain collector.

/Table 5.

Table 5.

## Main features of developing cumulus investigated in Flight 3.

Cloud Traverses

| Height feet | Mean Temp. $^{\circ}$F. | Turbulence | | Icing g/cu.m. | | Water Content g/cu.m. | | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|---|
| | | Max g | Mean g | Max | Mean | Max | Mean | | | |
| 15,000 | 17 | .3 | .2 | Not recorded | | .6 | .5 | 1 | 1 | Mixed ( crystals<br>( droplets<br>light rain |
| 13,000 | 22 | .3 | .2 | 1.2 | .8 | .6 | .4 | 3 | 1 | Mixed ( crystals<br>( snow<br>( hail<br>( droplets |
| 11,000 | 28 | .3 | .2 | Not recorded | | .7 | .3 | 6 | 2 | Mixed ( droplets<br>( rain<br>( snow |
| 9,000 | 36 | .5 | .3 | Nil | | 1.0 | .6 | 9 | 2 | droplets<br>rain |
| 7,000 | 40 | .3 | .2 | Nil | | .7 | .5 | 6 | 2 | droplets<br>rain |
| 5,000 | 47 | .3 | .2 | Nil | | .7 | .5 | 9 | 2 | droplets<br>rain |
| 3,000 | 53 | .2 | .2 | Nil | | .4 | .2 | 4 | 1 | droplets<br>rain |

Table 6

## PROJECT CU - FLIGHT 3

### Composition of cumulus cloud

| Height feet | Temp. $^{\circ}$F. | Water Droplets | Ice Crystals |
|---|---|---|---|
| 15,000 | 21 | Mainly 5-10μ, occasionally 140-240 μ. | Mainly a complicated dendritic structure in aggregates up to 170 μ across. Some droplets, included in structure, of 10-40μ diameter. |
| 13,000 | 23 | Mainly 14-40 μ, occasionally 5 and 80μ. Density variable between 150 and 700/cm$^3$. | Tetragonal and hexagonal plates of 20-170μ in aggregates up to 700 μ. Some single crystals with droplets attached. Separate droplets of 10-55μ. |
| 11,000 | 29 | Mainly 7-10 μ but occasionally 7-20μ. Fair number of 200-240 μ and apparently some snow. | None |

/9,000

Table 6 continued.

| Height feet | Temp. °F. | Water Droplets | Ice Crystals |
|---|---|---|---|
| 9,000 | About 36 | Mainly 5-10 μ, occasionally 60-80 μ and a few 350 μ and one 1100 μ. | None |
| 7,000 | About 40 | Mainly 10-20 μ but some areas of 20-30 μ. Occasionally 60-110 μ. Probably 700/cm$^3$. | None |
| 5,000 | 48 | Mainly 5-10 μ but some areas of 20-40 μ diameter. A few 120-380 μ. Density variable between 500 and 700/cm$^3$. | None |
| 3,000 | 54 | Mainly 14-40 μ and a few 100-350 μ diameter. Density less than 100/cm$^3$. | None |

Table 7

## PROJECT CU - FLIGHT 3.

### Composition of altostratus cloud

| Height feet | Temp. °F. | Water Droplets | Ice Crystals |
|---|---|---|---|
| 23,000 | -15 | Nil | Some tetragonal and a few hexagonal plates, but mainly hexagonal columns or a tending to columnar development. Plates mostly 170 μ diameter but some 20-100 μ. Columns 210 μ long x 70 μ or less broad. Few double or treble crystals, no aggregates. |
| 22,000 | -10 | Nil | Mainly hexagonal plates with a few tetragonal plates and hexagonal columns. Plates mainly 20-60 μ with a few 90 μ diameter stellar tendencies, some tetragons aggregated in dendritic forms. |
| 21,000 | -6 | Nil | Tetragonal plates and short columns. Plates 20-240 μ diameter, columns 100-170 x 70-100 μ, but mainly 200 x 80 μ. Often three or four columns joined by what appears to be a frozen droplet. |
| 20,000 | -2 | Occasionally 14 μ diameter. | Hexagonal and tetragonal plates, hexagonal predominating. A few short columns 100 x 70 μ. Plates 20-350 μ diameter but mainly about 170 μ. Many aggregates up to 1400 μ across and chains of 500 μ long. Plates show internal ribbing. |

/Flight 4.

Flight 4. 8/8/52.

1. Execution of the Flight. The investigation took place between 1100 and and 1430Z. After a sounding up to 10,000 ft. ICAN had been made over Cranfield a developing cumulus about 50 miles NNE of Dunstable was selected for investigation and traverses made at 8,000 ft. (above cloud tops) 6,500, 5,500, 5,000, 4,500 and 4,000 ft. (in cloud).

2. The soundings and their relation to observed cloud features. From the ascent which is shown on Fig. 6 it appears that:-

(a) From the parcel method, a surface temperature of 67°F and a dewpoint of 54°F would lead to a cloud base of about 900 mb. which agrees quite well with the observations made.

(b) Cloud tops however might have been expected to extend to 760 mb. In fact the tops were observed to be at about 810 mb. which coincides with a small layer of decreased lapse rate.

(c) The increase of temperature of the ambient air at 810 mb. required to stop vertical motion there is less than 1°F. This observed height of cloud tops might be accounted for by quite a small amount of entrainment or mixing around and just below this level where it is comparatively dry.

(d) Photographs of the clouds investigated are given in Figs. 7a and 7b.

3. Results of the traverses. The main cloud features are summarised in Table 8 and short notes are given on them below:-

(a) Cloud structure. A complex of clouds was investigated which consisted of two main clouds each 1-2 miles wide and 3,000 ft. deep and having a small cloud of the same width and 1,000 ft. deep to the south of it. In each case it appeared that the subsidiary clouds, situated in or closely associated with descending air, were the products of decay of old cells.

The most northerly (downwind) cloud was remarkable in that it appeared to be set in an ascending environment at all heights, within downdraughts occurring within the cloud itself. Rain was observed in the absence of ice crystals (the lowest temperature recorded in the cloud was +34.2°F). Presumably ice crystals could be present in the atmosphere above the freezing level but as this cloud was everywhere below it any rain within the cloud must have been due to another cause - probably coalescence. Rain was also observed in the southerly cloud where it appeared that the coalescence had advanced further.

(b) Composition. Droplets were sampled at all heights, and the maximum droplet diameter at a given level increased from 200 μ near the base of the most northerly cloud to 400μ (a small raindrop) near tops. In this cloud the droplet spectrum appeared to decrease in width with height as one would expect in the early stages of a simple coalescence process, though the model may be complicated, in fact, by the advance in time as the measurements were made. In the southernmost major cloud rain was also found, the diameter decreasing from 700 μ near base to 140 μ near tops. The spectrum of droplet sizes was relatively constant, sizes ranging usually from 7 μ to 55 μ. Details of the droplet sampling and the droplet distributions are given in Table 9; Table 10 and Fig. 8 give detailed distributions for the samples taken at 4,500 ft., 5,000 ft. and in rain close to cloud base.

Coalescence was almost certainly the rain making process again in this latter cloud but the model is probably not so simple as in the previous case. Whereas previously one could envisage a process in which the larger droplets were swept up by the large drops found at the base, thus reducing the basic droplet spectrum width with accompanying increase in rain drop size and spectrum width, in this case it is reasonable to suppose that this process has already taken place and that sedimentation has reversed the trend in maximum drop diameter with height. The constancy of droplet spectrum width is a disturbing feature but this may well be ascribed to the ascent of fresh bubbles into an already mature cloud.

The most noteworthy feature of the whole investigation was the unusual width of the droplet spectrum and this must be the feature most responsible for the coalescence rain production. The air mass concerned was of maritime origin and it may be that the larger droplets were formed on giant salt nuclei as proposed by Ludlam [5]. The great variability of the results of successive samplings within short distances is also notable.

(c) Vertical motions and turbulence. Vertical motions were usually small in the northern cell and never exceeded 15 ft./sec. when averaged over half a mile. The greatest velocities were observed at the lowest levels, i.e. late in time. In this cell rain was observed in descending air.

In the southern cloud velocities were generally less than 5 ft./sec. and were normally greater in the more southern regions. Here rain was found in ascending air and downdraughts were observed in its environment.

Traverses made below cloud indicated vertical motions of 1-2 ft./sec. but the general motion lacked coherence.

Turbulence was infrequent and small in both clouds and usually limited to the shear zones between vertical motions of opposing sense.

(d) Water contents were materially higher in the southern cloud than in the northern, values being generally about 0.5 g/m$^3$ in the one case and about 0.3 g/m$^3$ in the other.

Table 8.

Main cloud features of cumulus clouds investigated in Flight 4.

Cloud Traverses

| Height feet | Mean Temp. °F. | Turbulence | | Icing | | Water Content | | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|---|
| | | Max g | Mean g | Max g/cu.m. | Mean g/cu.m. | Max g/cu.m. | Mean g/cu.m. | | | |
| 8,000 | 35 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | Over top of cloud in clear air. |
| 6,500 | 40 | .2 | .1 | Nil | | .6 | .3 | 2 | 2 | Droplets<br>Small rain drops. |
| 5,500 | 44 | 0 | 0 | Nil | | .7 | .3 | 2.5 | 2 | Droplets<br>Small rain drops |
| 4,500 | 46 | .2 | .1 | Nil | | .4 | .2 | 5 | 5 (2 small) | Droplets<br>Rain drops. |

/Later

Table 8 continued.

Cloud Traverses

| Height feet | Mean Temp. °F. | Turbulence Max g | Turbulence Mean g | Icing Max Mean g/cu.m. | Water Content Max g/cu.m. | Water Content Mean g/cu.m. | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|
| Later runs | | | | | | | | | |
| 4,000 | 48 | .2 | 0 | Nil | - | - | 4 | 3 | Droplets |
| 5,000 | 45 | 0 | 0 | Nil | - | - | 4 | 3 | Droplets |

Table 9

### PROJECT CU - FLIGHT 4.

Composition of a developing cumulus cloud

| Height feet | Temp. °F. | Water Droplets | Ice Crystals |
|---|---|---|---|
| 6,500 | 42 | Occasionally 100-400 μ but mainly 14-20 μ diameter. Density 3-400/cm$^3$. | Nil |
| 5,500 | 44 | Mainly 7-55 μ diameter, occasionally 80-230 μ. Density $\gg$ 500/cm$^3$, occasionally diminishing to 3-400/cm$^3$. | Nil |
| 4,500 | 47 | 7-40 μ diameter, with some rain drops of up to 700 μ. Density 300/cm$^3$. | Nil |
| 4,000 | 48 | 7-40 μ diameter, but mainly 15-20 μ. A background of very small droplets $\leqslant$ 4 μ. Density about 100/cm$^3$. | Nil |
| 5,000 | 46 | 7-20 μ but mainly 10-15 μ diameter. Density 500/cm$^3$. | Nil |

/Table 10

Table 10

Flight 4. 8/8/52.

Droplet distribution by volume

| Droplet diameter "d" | Percentage of total by volume contained in droplets of diameter less than "d" | | | | | Height feet |
|---|---|---|---|---|---|---|
| | Sample 1 | 2 | 3 | 4 | 5 | |
| µ | | | | | | |
| 11 | 5.6 | 0.7 | | | 1.1 | 4,500 |
| 17 | 5.6 | 0.7 | | | 1.1 | |
| 25 | 95.8 | 24.6 | | | 70.6 | |
| 35 | 100.0 | 34.5 | | | 93.3 | |
| 50 | | 53.9 | | | 100.0 | |
| 70 | | 92.8 | | | | |
| 90 | | 100.0 | | | | |
| Other sizes | | 1,700µ | | | | |
| 11 | 0.8 | 3.7 | 33.8 | 24.2 | | 5,000 |
| 17 | 0.8 | 3.7 | 33.8 | 24.2 | | |
| 25 | 66.5 | 82.9 | 100.0 | 100.0 | | |
| 35 | 91.2 | 88.0 | | | 2.8 | |
| 50 | 95.9 | 100.0 | | | 6.3 | |
| 70 | 100.0 | | | | 14.2 | |
| 90 | | | | | 25.8 | |
| 110 | | | | | 32.1 | |
| 130 | | | | | 42.9 | |
| 150 | | | | | 53.2 | |
| 180 | | | | | 64.2 | |
| 220 | | | | | 100.0 | |
| 11 | | 0.1 | 0.1 | 0.4 | | Rain below base |
| 17 | | 0.1 | 0.1 | 0.4 | | |
| 25 | 1.0 | 0.7 | 3.3 | 10.1 | 1.1 | |
| 35 | 2.3 | 7.9 | 6.3 | 18.3 | 5.4 | |
| 50 | 4.0 | 14.3 | 8.8 | 28.3 | 9.6 | |
| 70 | 4.0 | 16.3 | 30.0 | 46.3 | 18.5 | |
| 90 | 20.3 | 21.3 | 82.2 | 71.6 | 23.4 | |
| 110 | 22.0 | 38.3 | 91.5 | 82.4 | 33.4 | |
| 130 | 31.5 | 38.3 | 100.0 | 100.0 | 50.6 | |
| 150 | 46.1 | 38.3 | | | 58.5 | |
| 180 | 62.7 | 48.3 | | | 64.5 | |
| 220 | 62.7 | 67.5 | | | 70.3 | |
| 260 | 62.7 | 100.0 | | | 100.0 | |
| 300 | 100.0 | | | | | |

Flight 5. 12/8/52.

1. Execution of the Flight. The cloud on this occasion was mainly of a stratocumulus rather than a cumulus type, although it did contain some developing cumulus. Bases of the layer were generally 2,500 ft. and tops generally 3,700 ft.

Traverses were made below cloud at 700, 1,200, 1,700 and 2,200 ft. An ascent was made from 900 to 8,100 ft. during which observations were made every 200 ft.

Unfortunately records of the ultra-rapid thermometer which were made during these traverses were lost due to a camera fault; a certain amount of information is available from the readings of the flat plate thermometer, hygrometer, height and airspeed and visual observations.

One traverse was made penetrating cloud tops and measurements of water content and droplet size obtained.

2. The sounding and its relation to observed features. The aircraft sounding - including winds from 1400Z Larkhill radio sonde - from 900 to 8,100 ft. is plotted on Fig. 9. Estimated surface temperatures and humidities for the Cranfield area at this time are also given.

Considering Fig. 9 it appears that:-

(a) Lowest cloud base was 925 mb. which is consistent with parcel method calculation based on a surface temperature of 69°F and surface dewpoint of 59°F. The actual values were:- Temperature 75°F, dewpoint 53°F.

(b) With this condensation level and ascent along a saturated adiabatic cloud top might be expected to be 860 mb. but was actually 885 mb.

(c) As there was a very dry layer also commencing at 885 mb. it appears likely that entrainment at this level was effective in limiting the cloud tops.

Drizzle was observed beneath cloud but was not actually observed to reach the surface.

3. The horizontal traverses above and below cloud. Summaries of the various traverses follow:-

(i) 2,200 ft. 33 miles under stratocumulus thinning after 12 miles, occasional breaks. A developing cumulus breaking through the sheet after 22 miles. Drizzle after 8 miles.

Updraughts and downdraughts were mainly random, about half a mile to a mile across and having strength 1-2 ft./sec. though exceptionally 3-5 ft./sec. at the beginning. There was no correlation with particular cloud features though a downdraught of 8 ft./sec. was associated with the drizzle.

Temperature rose unsteadily 1.8°F throughout the run; the gradient was normally about 0.2°F/mile but occasionally rose to 0.8 - 1.2°F/mile. There was a tendency for higher temperature to be associated with updraughts and lower to be associated with downdraughts in the case of the stronger currents.

Frost point varied between 54° and 57°F throughout the run, being lowest and most variable during the last half of the run when under broken cloud.

Turbulence was very slight rising to an estimated value of 0.1g only occasionally.

(ii) 1,700 ft. 22 miles under blue sky with only occasional wisps of cloud. Up and down currents were random and usually half a mile to a mile across and 1-2 ft./sec. in strength. A downdraught of 4 ft./sec. was observed near the middle of the run.

Temperature varied about 1.3°F throughout the run, usually at the rate of 0.2 - 0.4°F/mile though occasionally 1.6°F/mile. There was no strong correlation between direction of vertical motion and temperature differences though a strong downdraught was associated with a rapidly falling temperature.

Frost point was very steady at about +55°F.

Turbulence was estimated to be about 0.1g and was experienced over most of the run.

(iii) 1,200 ft. 17 miles half under blue sky, half under broken cloud. Up and down currents were random and usually half a mile to a mile wide and of speed 1-2 ft./sec. under blue sky. Under the broken cloud an upcurrent of speed $>$ 2 ft./sec. and width about 2 miles was experienced. At the edge of the cloud sheet there was turbulent motion in which the vertical motion changed from $>$ 11 ft./sec. upwards to $>$ 8 ft./sec. downwards.

Temperature varied about 0.8°F between extreme values throughout, and was higher and more variable in the blue sky region. Temperature gradients were usually 0.2°F/mile but occasionally 1.0°F/mile. There was some correlation between direction of vertical motion and sense of temperature change beneath blue sky, but not beneath cloud.

Turbulence was estimated to be 0.1g and was rare, being experienced as broken cloud was approached.

(iv) 700 ft. 17 miles under wisps and broken cloud, cloud predominating in the first two thirds of the run.

Up and down currents were random with a slight tendency for updraughts to be associated with wisps and downdraughts to be associated with breaks. Strength was normally 1-2 ft./sec. and width about half a mile.

Temperature varied about 1.3°F between extremes over the run, reaching a maximum value just before the cloud edge encountered after two thirds of the run. Gradients were generally 0.8 - 1.4°F/mile. There was no correlation between temperature and vertical motion.

Frost point was steady at 57°F.

Turbulence was estimated to be 0.1 - 0.2g.

(v) 4,000 ft. 25 miles entirely above cloud with three short penetrations into cloud.

Up and down currents were rather more coherent and less frequent than below cloud, and in general linked with the tops penetrated. Width was generally 1-2 miles and strength variable between 2 and 15 ft./sec., the highest value being observed in downdraught. Drizzle was observed in one top in an updraught.

Temperature, when corrected for changes of height, varied about 2°F between extremes during the run, tending to rise 0.6 to 0.9°F in cloud, although instrumental uncertainties make this last conclusion doubtful.

No turbulence was recorded.

4. Data obtained in cloud tops. Water contents were 0.5 $g/m^3$ in the updraught portions of cloud and 0.4 $g/m^3$ in the downdraught. A value of 0.25 $g/m^3$ was recorded in drizzle-bearing wisp of cloud. Cloud base temperature was about 57°F and cloud top temperature 51°.

Droplet diameters varied between 8 and 240 μ. The width of spectrum was quite variable ranging from 8 to 30 μ to 8 to 240 μ at the same height.

The water contents are less than a half of that anticipated from a saturated adiabatic ascent which tends to confirm the entrainment ideas suggested above.

The largest water drops are drizzle size and are often observed in clouds similar to those explored on this occasion. The large droplets found in this cloud must have been formed by a coalescence process which had been going on for some considerable time. This contention is supported by the fact that the cloud sheet had been in existence for about forty-eight hours.

Flight 6. 13/8/52.

1. Execution of Flight. Horizontal runs were made at 500, 1,000, 2,000, 3,000 and 4,000 ft. partly beneath a sheet of stratocumulus, occasionally with mammato-form base, and partly under cumulus. This sheet of stratocumulus, which occupied the quadrant north west of Cranfield, had a base level at 7,700 ft. which was rather above that of the surrounding cumulus. Beneath it there was decaying cumulus which occasionally grew into and through the sheet. Some rain was observed below the stratocumulus. An ascent measuring temperature and frost points was then made to 12,000 ft.

On conclusion of the clear air work, runs were made in cloud tops at 10,000, and also at 9,000 and 8,000 ft. whilst on track for base.

2. The soundings and their relation to observed cloud features. The ascent is plotted on Fig. 10 and the following notes are given:-

(a) From parcel method considerations, a surface temperature of 74°F and a dewpoint of 51°F would have been necessary to produce a condensation level at 830 mb. Unfortunately we have no good observation of the height of the cumulus base. The stratocumulus base which was somewhat higher was at about 750 mb.

(b) With this condensation level cumulus tops might have been expected to extend to 740 mb. and occasionally 650 mb. There was only a small "stop" at the former but at the latter there was a well marked inversion and a considerable decrease in the frost point. Tops were actually at 680 to 690 mb., just below the inversion.

(c) Surface temperatures in the Cranfield area may not quite have attained the value of 74°F as the reported 1200 and 1500 hours temperatures are 70°F. This would be consistent with the fact that the cumulus clouds were rather sporadic and tending to decay during the afternoon and the stratocumulus layer may have been mainly the result of a small zone of wind shear between 800 and 750 mb.

(d) Photographs of the clouds are given in Figs. 11 a and b.

3. Results of the traverses below cloud. These will not be described in detail as they were very similar in execution to those made in Flight 5. Each run was about 8 miles in length. The main features are summarised below:-

(i) Rain was encountered both at 4,000 and 3,000 ft. It was associated with well marked downdraughts 1 to 2 miles across on each occasion.

(ii) Elsewhere up and downdraughts were small probably almost always less than 5 ft./sec. and usually randomly distributed. Their extent varied from ½ to 2 miles across. At 4,000 ft. and 3,000 ft. the downdraughts were mainly on the south side and the updraughts on the north side of the clouds while under clear skies there were usually small downdraughts. Downdraughts were noticed under the mammato base of the stratocumulus cloud.

(iii) Temperature variations were about 0.5°F in a half a mile and the excess in updraughts was about 0.3°F, the deficiency in downdraughts being generally smaller. At 2,000 ft. and below there was no apparent correlation between temperature variation and the direction of vertical motions.

(iv) Turbulence was generally less than 0.2g and became less probable with increase in height. At 3,000 ft. turbulence was confined to the edges of cloud and at 4,000 ft. no turbulence greater than 0.1g was observed.

/4.

4. Run above cloud at 11,000 ft. This was above cloud tops and mainly in a downdraught area where it was noted that temperature was about 1.5°F less than in corresponding updraught areas.

5. Runs in cloud. The main features are given in Table No. 11 and the following additional remarks are given:-

(a) Cloud structure. The clouds investigated were predominantly stratocumulus with only occasional cumulus. Most of the phenomena appeared to vary systematically over distances of 2-3 miles in the lower portion of the cloud and 1-1½ miles in the upper portion. Rain was encountered at all heights and was heaviest in the tops.

(b) Composition. Droplet samples were obtained at all heights and details are given in Table 12. In Table 13 and in Fig. 12 results of successive samplings at 200 yd. intervals are shown. Photographs are also shown in Figs. 13 a and b of the type of samples obtained. The droplet spectrum appeared to vary irregularly throughout the cloud while the drizzle or raindrop size increased towards the base. No crystals were found and there was no icing although cloud tops were slightly higher than the freezing level.

There was higher cloud in the area which might possibly have provided a source of ice nuclei but it appears more likely that coalescence was the principal factor in producing drops of rain size from this cloud. The droplet spectra tend to support this view, and it appears that whereas in some parts of the cloud the coalescence process was well advanced in others it was in an early stage.

(c) Vertical motions and turbulence. Up and downdraughts were ill defined both near the base and the tops of this cloud but in the central portions appeared occasionally to exceed 10 ft./sec.

Turbulence was generally between 0.1g and 0.2g and was frequent near the base of this cloud.

(d) Water content and icing. As cloud top temperatures were about 29°F kinetic heating effects would have been sufficient to prevent any icing.

Water content was very variable at all levels and reached peak values of 0.5 g/m$^3$ both at the tops and in the base of the cloud. It appeared to be rather less in the middle. It was noticed that the highest water content measurements tended to coincide with the rain areas.

Table No. 11

Composition of stratocumulus cloud investigated in Flight 6.

Cloud Traverses

| Height feet | Mean Temp. °F. | Turbulence Max g | Turbulence Mean g | Icing Max Mean g/cu.m. | Water Content Max g/cu.m. | Water Content Mean g/cu.m. | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|
| 10,000 | 29 | .2 | .1 | Nil | .5 | .2 | 8 | Several (Cu. and St.Cu.) | Droplets and Rain. |
| 9,000 | 30 | .2 | .1 | Nil | .3 | .1 | 8 | Several (Cu. and St.Cu.) | Droplets<br>Rain. |
| 8,000 | 35 | .2 | .1 | Nil | .5 | .2 | 8 | Several (Cu. and St.Cu.) | Droplets<br>Rain |

Table 12

## PROJECT CU - FLIGHT 6

### Composition of stratocumulus cloud

| Height feet | Temp. °F. | Water Droplets | Ice Crystals |
|---|---|---|---|
| 10,000 | 28 | Mainly 10-20 μ diameter but some 30-40 μ and a few 100-350 μ. Density 200-300/cm³. | None |
| 9,000 | 30 | Early samples: Mainly 10-55 μ diameter with greater concentration around 10 μ and 55 μ. A few 350 μ. Density 200-300/cm³.<br>Later samples:<br>Mainly 7-20 μ with greatest concentration at 10 μ. Occasionally 60 μ. Density 300-400/cm³. | None |
| 8,000 | 35 | Early samples:<br>Mainly 10-40 μ but some 350-520 μ. Density 400/cm³.<br>Later samples:<br>Mainly 14-40 μ with some 70-220 μ. Density 100/cm³. | None |

Table 13

Flight 6. 13/8/52.

### Droplet distribution by volume

| Droplet diameter "d" | Percentage of total by volume contained in droplets of diameter less than "d" | | | | | Height feet |
|---|---|---|---|---|---|---|
| | Sample 1 | 2 | 3 | 4 | 5 | |
| μ | | | | | | |
| 11 | 0.3 | 0.1 | 0.1 | | 0.1 | 8,000 |
| 17 | 0.3 | 0.1 | 0.1 | | 0.1 | |
| 25 | 3.1 | 8.4 | 7.6 | 0.9 | 4.9 | |
| 35 | 17.7 | 13.1 | 20.0 | 3.2 | 21.9 | |
| 50 | 37.0 | 20.0 | 30.6 | 8.7 | 35.9 | |
| 70 | 83.7 | 48.6 | 48.7 | 30.4 | 71.0 | |
| 90 | 100.0 | 61.0 | 59.4 | 48.0 | 87.0 | |
| 110 | | 79.1 | 59.4 | 58.3 | 87.0 | |
| 130 | | 100.0 | 66.6 | 78.9 | 100.0 | |
| 150 | | | 66.6 | 78.9 | | |
| 180 | | | 100.0 | 90.0 | | |
| 220 | | | | 100.0 | | |

/Flight 7.

## Flight 7. 14/8/52

1. Execution of the Flight. An ascent was made to 20,000 ft. ICAN in the Mosquito aircraft in the morning and further measurements were made with the Hastings aircraft in the afternoon.

After the Hastings sounding of temperature and humidity over Cranfield from 1,000 to 9,000 ft. it was decided, in the absence of any large cumulus, to concentrate on clear air runs below cloud base level; owing to the recording camera fault previously mentioned much of the information from the ultra-rapid thermometer was lost.

Horizontal runs were made at 9,000, 8,000, 6,500, 6,000, 5,500, 4,000, 2,000, 1,000 and 500 ft. ICAN. Cloud base in the afternoon was 5,300 ft. and tops 8,000 ft. and data on conditions within the cloud were obtained on the 8,000 and 6,500 ft. runs, the 6,000 ft. run being entirely in clear air.

2. The soundings and their relation to observed cloud features. The results of the 1100Z Mosquito ascent and the 1500Z Hastings ascent are plotted on Figs. 14 and 15. Considering these diagrams:-

(a) (i) From parcel method calculations the lowest condensation level at 1100Z would be about 840 mb. (actual value 850 mb.) with a surface temperature of about 68°F and dewpoint of 47°F. Observed surface values at this time were about 68°F and 53°F. The water content line above 950 mb. suggests that parcels originate mainly in the layer beneath 950 mb.

(ii) Similar calculations show that the lowest condensation level at 1500Z would be at about 815 mb. (actual value 840 mb.) with a surface dewpoint of about 45°F and temperature of 70°F. Observed surface values at this time were 69°F temperature and 55°F dewpoint.

(b) (i) There was an isothermal layer between 780 mb. and about 720 mb. at 1100Z which effectively limited penetration by parcels to about 760 mb. a value in good agreement with the observed tops at 745 mb.

(ii) The ascent made at 1500Z indicates a small inversion - of 3°F - between 750 and 720 mb. which would limit tops to about 740 mb. - the observed tops being at about 750 mb.

(c) (i) The 1100Z ascent indicates a marked drying of the air above 870 mb. which indicates that entrainment cannot have been a significant feature in development at this stage.

(ii) The 1500Z ascent is quite moist to 750 mb. above which there is pronounced drying. The change since the morning must presumably be ascribed to convective transport of water vapour upwards. Even so, surface dewpoints are surprisingly high compared to those measured at the lowest point in the aircraft ascent. Normally when there is considerable convection taking place only a very small hydro-lapse would be expected if there were much mixing with the environment. In this case it must be concluded that any air taking part in cloud formation must have originated below 950 mb. in the morning and below 980 mb. in the afternoon - both these pressures being relative to a corrected surface pressure of 1013 mb.

3. Runs below cloud. These runs, which were made at heights of 4,000, 2,000, 1,000 and 500 ft., showed features very similar to those described in connection with Flights 5 and 6.

(i) Vertical motions were small but occasionally speeds of about 5 ft./sec. were measured. These up and downdraughts were more organised below 2,000 ft. than above and were then between 1 and $2\frac{1}{2}$ miles in extent. Higher up they were more randomly distributed and no size can be given. Small downdraughts again usually occurred beneath clear sky while both up and downdraughts were noted below cloud.

(ii) The temperature varied by about 0.5°F in half a mile but there seemed to be no correlation between the regions of higher or lower temperature and the direction of the vertical motions.

(iii) Turbulence at all levels was usually less than 0.1g but bumps of 0.2g were occasionally reported during the 2,000 and 4,000 ft. runs, usually near the edges of the up or downdraughts.

Run above cloud at 9,000 ft. Vertical motions were irregular and small and no bumpiness was noted. Temperature varied with distance by about the same amount as below cloud and seemed to have little correlation with the direction of the vertical motions.

Run in cloud. Short notes are given as the runs were carried out at 8,000 and 6,500 ft. only.

(a) Cloud structure. The cloud consisted of mixed small cumulus and stratocumulus about 3,000 ft. thick and presented no regular structure. Cloud tops were just below freezing level.

(b) Composition. The cloud consisted of water droplets some of which reached drizzle size. The droplet spectra were rather wider than usual and details of these are given in Table 14 and Fig. 16. The drizzle forming process in this cloud was almost certainly coalescence only. No rain or drizzle was observed outside the cloud. The droplet spectra again vary considerably in the horizontal which is presumably associated with different stages of the coalescence process in different cells.

(c) Vertical motions and turbulence. Up and downdraughts were small and random. Turbulence was infrequent and usually between 0.1 and 0.2g. At 8,000 ft. it was observed most at the southern edge of the clouds and at 5,500 ft. mostly at the northern edges.

(d) There were no measurements made of water content.

Table 14

Flight 7. 14/8/52.

Droplet distribution by volume

| Droplet diameter "d" | Percentage of total by volume contained in droplets of diameter less than "d". | | | | | Height feet |
|---|---|---|---|---|---|---|
| | Sample 1 | 2 | 3 | 4 | 5 | |
| μ | | | | | | |
| 11 | 0.2 | 0.1 | | | 0.1 | 8,000 |
| 17 | 5.0 | 7.8 | 1.2 | | 5.4 | |
| 25 | 9.4 | 19.3 | 8.5 | | 7.8 | |
| 35 | 19.7 | 44.0 | 33.2 | | 12.2 | |
| 50 | 49.5 | 51.9 | 78.3 | | 27.2 | |
| 70 | 68.8 | 69.8 | 86.9 | | 37.2 | |
| 90 | 72.7 | 69.8 | 100.0 | | 44.2 | |
| 110 | 79.3 | 100.0 | | | 56.4 | |
| 130 | 100.0 | | | | 64.0 | |
| 150 | | | | | 69.6 | |
| 180 | | | | | 80.7 | |
| 220 | | | | | 100.0 | |

/11

Table 14 continued

| Droplet diameter "d" | Percentage of total by volume contained in droplets of diameter less than "d" | | | | | Height feet |
|---|---|---|---|---|---|---|
| | Sample 1 | 2 | 3 | 4 | 5 | |
| μ | | | | | | |
| 11 | 0.1 | | 0.1 | | 0.1 | 6,500 |
| 17 | 0.1 | | 0.1 | | 0.1 | |
| 25 | 9.4 | 8.5 | 13.2 | | 4.1 | |
| 35 | 31.3 | 26.1 | 31.6 | | 15.1 | |
| 50 | 51.1 | 42.4 | 70.3 | | 37.7 | |
| 70 | 66.7 | 69.9 | 82.9 | | 87.8 | |
| 90 | 70.4 | 100.0 | 89.0 | | 91.9 | |
| 110 | 70.4 | | 93.6 | | 100.0 | |
| 130 | 70.4 | | 100.0 | | | |
| 150 | 70.4 | | | | | |
| 180 | 100.0 | | | | | |

Flight 8. 15/8/52.

1. Execution of the Flight. As the ground organisation of the project was being dispersed on this day, the more vigorous convection to the west - south-west of Farnborough was investigated. A cumulus congestus amongst several that had already become glaciated was selected for investigation after an aircraft sounding had been made to 20,000 ft. ICAN, but proved to be beyond the range of Farnborough Long Range Radar. Eventually a line of cumulus congestus, running east-west and comprising clouds of all sizes, was selected. Most of the cells were still growing and many had reached 15,000 - 20,000 ft. The growth cycle appeared to reach maturity in about 20 minutes.

Traverses were made through this cloud sheet on northerly and southerly headings at 2,000 ft. intervals from 19,000 to 9,000 ft. At the latter height the aircraft began to enter the radar ground echo of Farnborough Control and the investigation was abandoned. No clear air traverses below cloud were made.

The rotating disc icing meter was unserviceable on this flight and no quantitative icing observations are presented though airframe icing was reported at all levels down to 11,000 ft. in the cloud.

Cloud photographs are presented in Figs. 18a and b.

2. The soundings and their relation to observed cloud features. The results of the sounding have been plotted in Fig. 17. Considering this diagram:-

(a) From parcel method calculations the lowest condensation level would be at about 950 mb., rising later to 905 mb., relative to a corrected surface pressure of 1013 mb., with a surface temperature of about 68°F and dewpoint of 59°F.

The lowest condensation level was at about 945 mb. and this appears to have been a general value.

A second condensation level was observed at 820 mb. Just below this height there was considerable conditional instability and above there was moderate wind shear; both of these factors are presumably of importance in the formation of the stratocumulus layer observed between 820 and about 750 mb. Above 700 mb. there was considerable drying and a marked decrease in shear.

Although entrainment did not inhibit the growth of vigorous cumulus clouds the values of water content measured within them indicates that it must have been a factor of importance.

/(b)

(b) By similar calculations tops would be limited locally to 920 mb. with another stop at 670 mb. - this latter "stop" being reinforced by the marked drying at 700 mb. The marked instability throughout the ascent on this day would permit some cloud to develop far beyond 450 mb. in the absence of any great amount of entrainment.

Observed tops were all about 895 mb. for the lowest stratocumulus layer, about 750 mb. for the other stratocumulus layer, and generally 700 to 530 mb. for cumulus, although some clouds far to the south had anvils above 300 mb.

The fact that many cumulus developed into the very dry environment above 700 mb. indicates that again entrainment is of little importance in the higher portions of convective clouds at least especially where there is marked instability.

3. Results of the traverses. The main features are shown in Table 15 and short notes are also given below:-

(a) Cloud structure. The cloud investigated was the most easterly of a line of similar clouds lying east-west. The investigation comprised a series of traverses on northerly and southerly headings. A single cell was found at both upper and lower levels though two cells were found at 11,000 and 13,000 ft. Cell size was about 1-2 miles.

(b) Cloud composition. It is interesting that although this cloud had reached a temperature of +6°F in its upper regions, no ice crystals had appeared: it is usually expected that crystal formation would commence at temperatures of the order of +12°F. With large cumulonimbus clouds in the area there must have been a plentiful supply of ice crystal nuclei.

Although crystals had not appeared it was apparent that the cloud was approaching maturity since a downdraught was setting in in the upper levels.

The detailed droplet spectrum is presented in Table 16 and Fig. 19. It is notable that the spectrum varied considerably in the horizontal over distances as short as 200 yd. and also in the vertical. A few very large drops were reported at 17,000 and 15,000 ft. It is apparent from the spectra that the coalescence process was considerably advanced and that any rain production later would be due to a complicated mixture of crystal and coalescence processes.

(c) Vertical currents and turbulence. Although vertical motions over much of the cloud were less than 10 ft./sec. vertical motions of the order of 20 ft./sec. averaged over half a mile were by no means uncommon at all levels.

The strength of these vertical motions must presumably be associated with the large positive area evident on the tephigram of this ascent. To the strength of these currents may also be ascribed the persistence of supercooled drops at the abnormally low temperature previously mentioned.

The cloud and its environment appears to contain regularly arranged broad ascending and descending currents of air of lateral dimension 5 miles each having at its centre a core of lateral dimension about 2 miles moving in opposite sense to its environment. The cloud examined was set half in an ascending current and half in a descending one with the descending core of the broad ascending current beginning to encroach into the broad, descending current at the top of the cloud.

Turbulence was found normally in the shear zone between vertical motions of opposing sense. It was observed at all levels below 19,000 ft. and was usually greater than 0.2g. Turbulence estimated to be 0.7g - the highest observed during 1952 by the Meteorological Research Flight - was observed at both 15,000 and 13,000 ft.

(d) Water content. Water content was normally greater than 0.2g/m$^3$ and occasionally exceeded 1.6g/m$^3$ (the limitation of the instrument). It appeared to reach a minimum at 13,000 ft.

(e) Temperature in cloud. Temperature in cloud, though somewhat suspect for reasons previously mentioned, appeared to be quite variable, being higher in updraught and lower in downdraught regions. The temperature differential between these areas was often about 1-2°F and the maxima appeared to coincide with maxima in water content.

(f) Run above cloud at 19,000 ft. Temperature rose gradually from about -7°F to -3°F during this run, the lower temperatures apparently being associated with downdraught areas and the higher with updraughts.

Vertical motions were usually less than 5 ft./sec. but a downdraught was observed above cloud of about 20 ft./sec.

Table 15.

Main cloud features of cumulus congestus investigated in Flight 8.

Cloud Traverses

| Height feet | Mean Temp. °F. | Turbulence | | Icing | Water Content | | Extent miles | No. of cells | Remarks - composition |
|---|---|---|---|---|---|---|---|---|---|
| | | Max g | Mean g | Max Mean g/cu.m. | Max g/cu.m. | Mean g/cu.m. | | | |
| 17,000 | +6 | .3 | .2 | Meter u/s but ice observed on aircraft. | >1.6 | - | 2 | 1 | Water droplets up to 80µ diameter. |
| 15,000 | +12 | .7 | .3 | " | >1.6 | .7 | 3 | 1 | Water droplets up to 120 µ diam. occasionally 200-300 µ. |
| 13,000 | +23 | .7 | .3 | " | .6 | .3 | 3 | 2 | Water droplets up to 60 µ diameter. |
| 11,000 | +27 | .4 | .2 | - | >1.6 | .3 | 4 | 2 | Water droplets up to 100 µ diameter. |
| 9,000 | +36 | .3 | .2 | - | 1.0 | .6 | 3 | 1 | Water droplets up to 100 µ diameter. |

Lower levels not investigated. Base was probably about 3,000 ft.

/Table 16

Table 16

Flight 8

Droplet distribution by volume

| Droplet diameter "d" | Percentage of total by volume contained in droplets of diameter less than "d" | | | | | Height feet |
|---|---|---|---|---|---|---|
| | Sample 1 | 2 | 3 | 4 | 5 | |
| μ | | | | | | |
| 11 | 0.1 | | | 0.1 | | 17,000 |
| 17 | 0.1 | | | 0.1 | | |
| 25 | 6.4 | | | 4.1 | | |
| 35 | 13.9 | | | 4.1 | | |
| 50 | 38.8 | | | 25.7 | | |
| 70 | 85.7 | | | 39.0 | | |
| 90 | 100.0 | | | 46.8 | | |
| 110 | | | | 46.8 | | |
| 130 | | | | 100.0 | | |
| Other sizes | | | 1 drop 500 μ | | | |
| 11 | 0.1 | 0.1 | | | | 15,000 |
| 17 | 0.1 | 0.1 | | | | |
| 25 | 3.0 | 1.6 | 0.4 | 2.5 | | |
| 35 | 5.0 | 5.3 | 0.4 | 2.5 | | |
| 50 | 13.0 | 17.9 | 2.1 | 12.2 | | |
| 70 | 30.1 | 47.3 | 4.1 | 21.6 | | |
| 90 | 57.0 | 66.3 | 7.3 | 43.8 | | |
| 110 | 57.0 | 88.6 | 7.3 | 43.8 | | |
| 130 | 57.0 | 100.0 | 10.0 | 100.0 | | |
| 150 | 57.0 | | 10.0 | | | |
| 180 | 100.0 | | 28.7 | | | |
| 220 | | | 28.7 | | | |
| 260 | | | 49.9 | | | |
| 300 | | | 100.0 | | | |
| Other sizes | 1 drop 1,300 μ | | | | 1 drop 500μ | |
| 11 | 0.2 | 0.1 | 0.4 | 0.1 | | 13,000 |
| 17 | 0.2 | 0.1 | 0.4 | 0.1 | | |
| 25 | 40.0 | 62.4 | 34.5 | 20.3 | | |
| 35 | 94.7 | 95.4 | 56.1 | 69.4 | | |
| 50 | 100.0 | 100.0 | 61.7 | 86.5 | | |
| 70 | | | 100.0 | 91.7 | | |
| 90 | | | | 91.7 | | |
| 110 | | | | 100.0 | | |
| 11 | 0.1 | | | | 0.5 | 11,000 |
| 17 | 0.1 | | | | 0.5 | |
| 25 | 20.5 | | | 6.3 | 5.7 | |
| 35 | 71.1 | | | 21.0 | 36.3 | |
| 50 | 71.9 | | | 37.4 | 66.6 | |
| 70 | 82.9 | | | 77.3 | 100.0 | |
| 90 | 82.9 | | | 87.6 | | |
| 110 | 100.0 | | | 100.0 | | |

/11

Table 16 continued

| Droplet diameter "d" | Percentage of total by volume contained in droplets of diameter less than "d" | | | | | Height feet |
|---|---|---|---|---|---|---|
| | Sample 1 | 2 | 3 | 4 | 5 | |
| μ | | | | | | |
| 11 | 1.1 | | | | | 9,000 |
| 17 | 1.1 | | | | | |
| 25 | 7.9 | | | 3.7 | | |
| 35 | 13.7 | | | 3.7 | | |
| 50 | 24.2 | | | 26.9 | | |
| 70 | 35.1 | | | 63.5 | | |
| 90 | 41.5 | | | 63.5 | | |
| 110 | 47.8 | | | 63.5 | | |
| 130 | 69.6 | | | 100.0 | | |
| 150 | 87.0 | | | | | |
| 180 | 87.0 | | | | | |
| 220 | 100.0 | | | | | |

M.R.P. 826

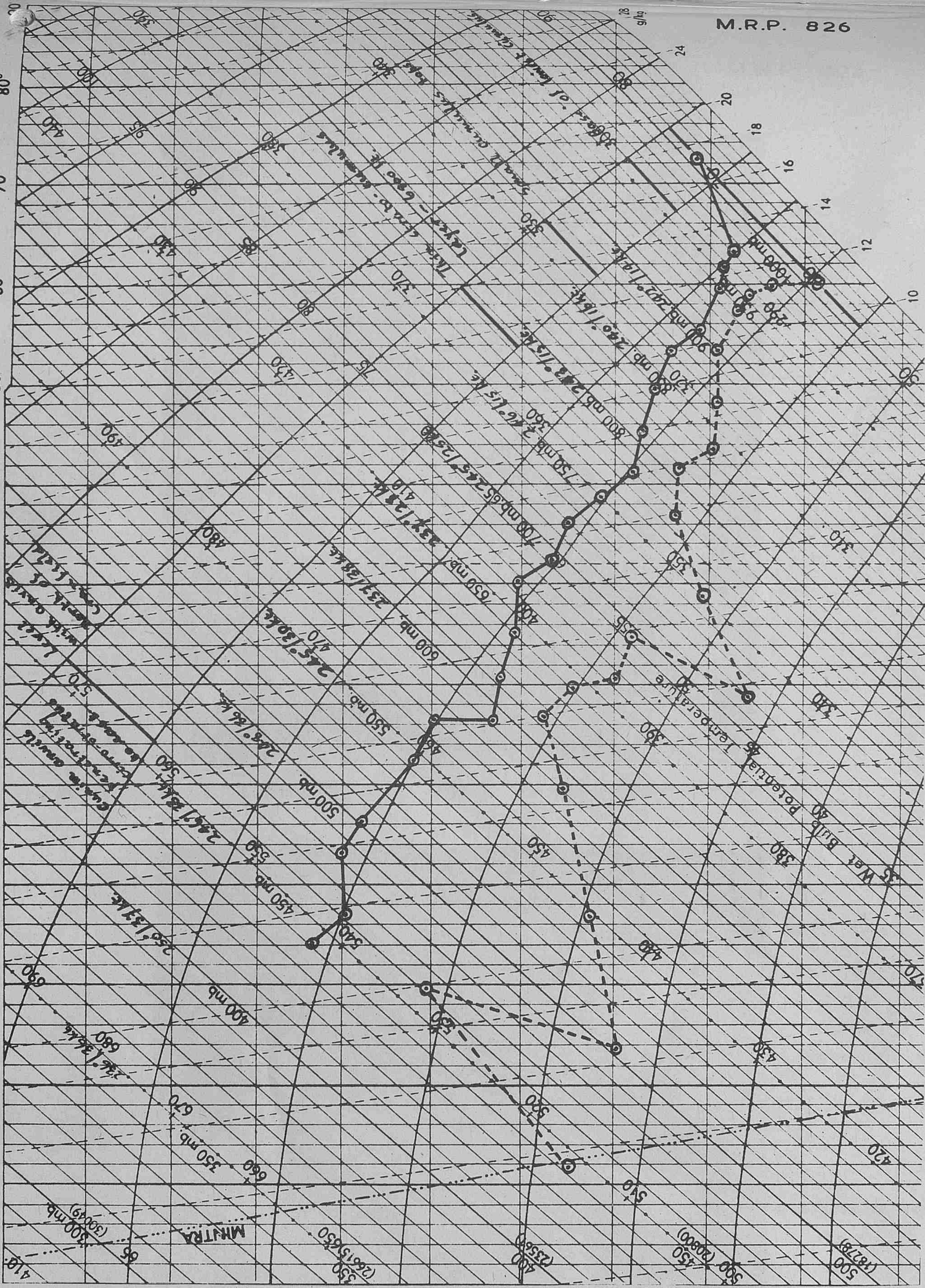


Fig. 1 Hastings ascent northwards from Cranfield on 5.8.'52

at about 16.00 Z.

Flight 1.    Figs. 2a and b

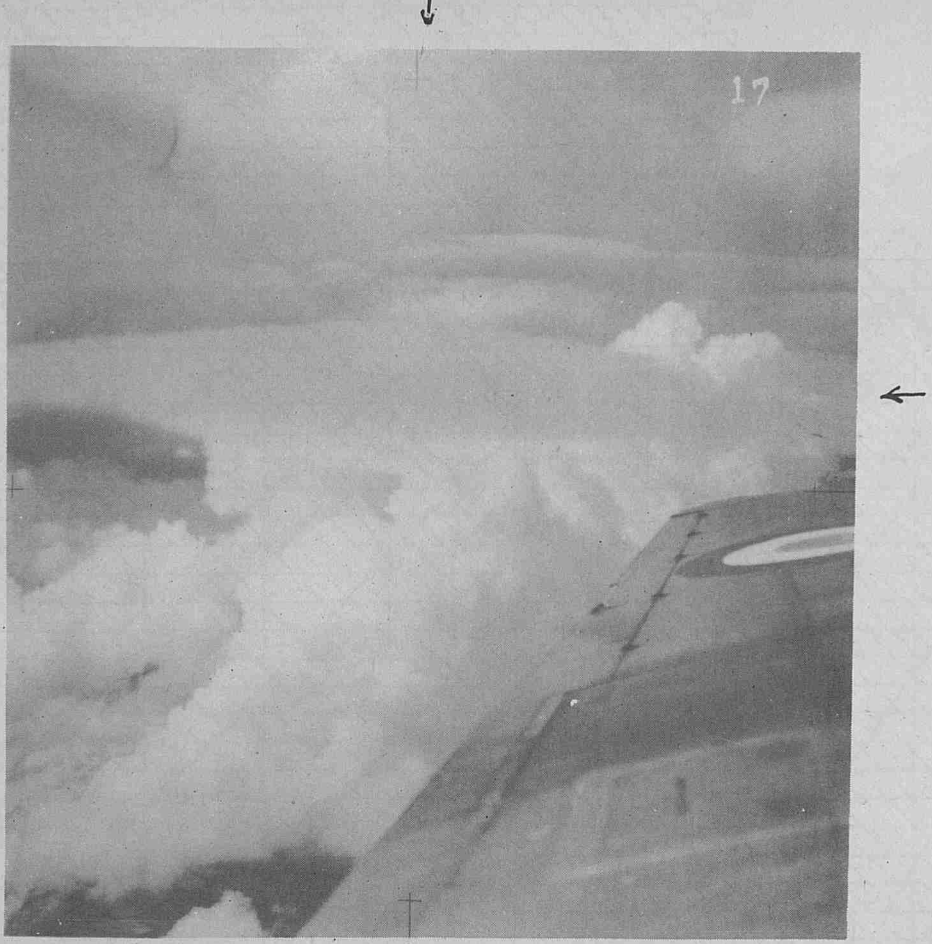


Fig. 2a  Lenticular strato-cumulus at 6800 ft. amongst lower cumulus tops.

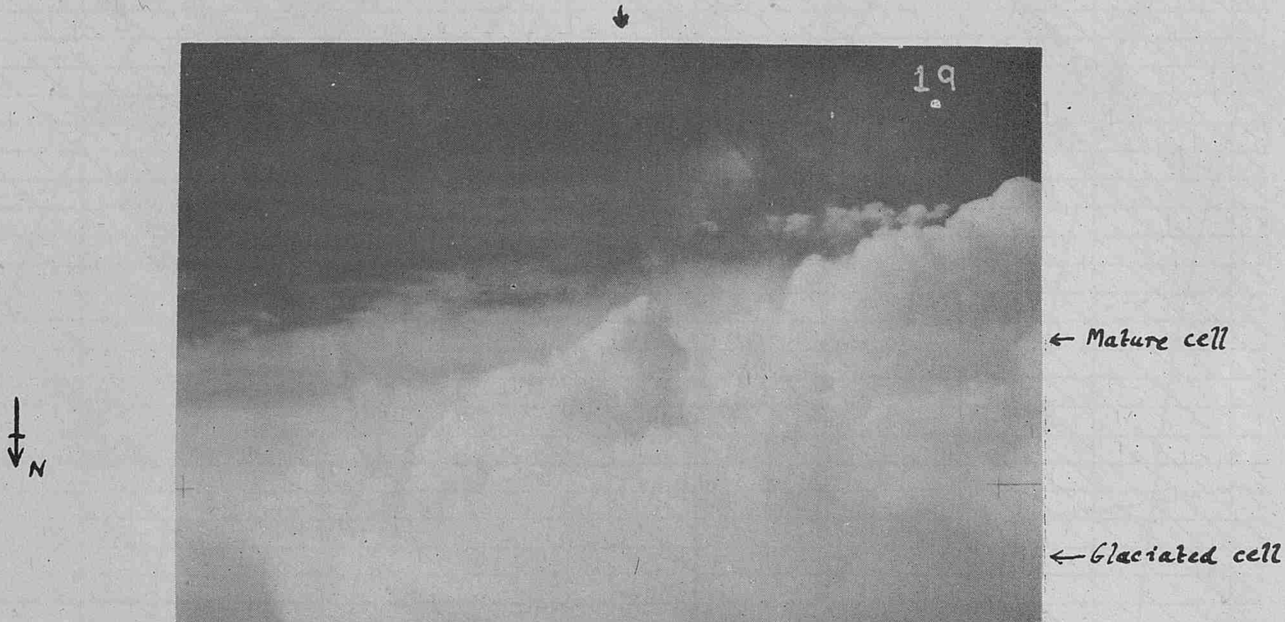

19
← Mature cell
← Glaciated cell
N

Fig. 2b  Cloud investigated, aircraft at 22000 ft. and approaching from North.

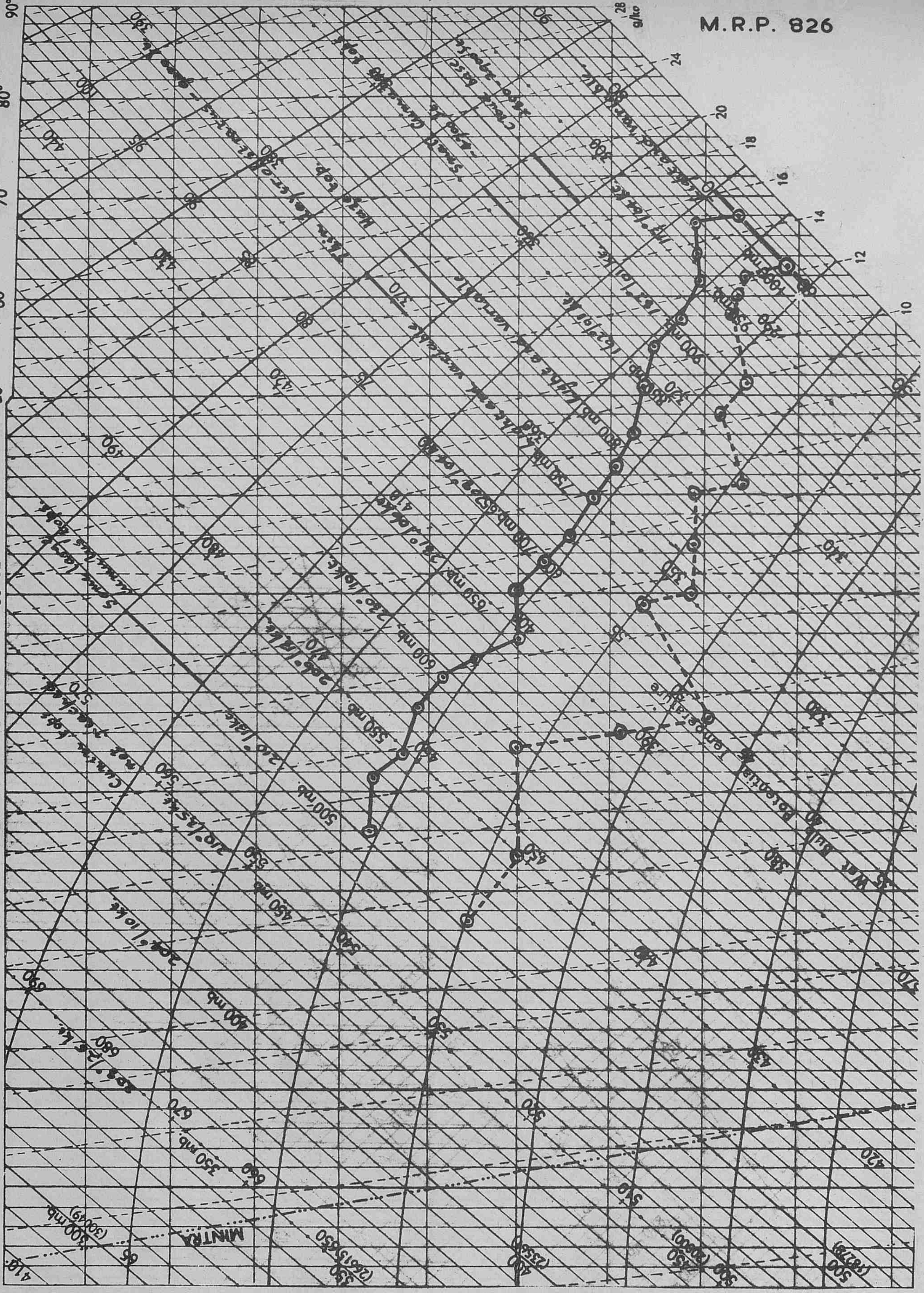


Fig. 3 Hastings ascent over Cranfield on 6·8·52 at about 14.00 Z

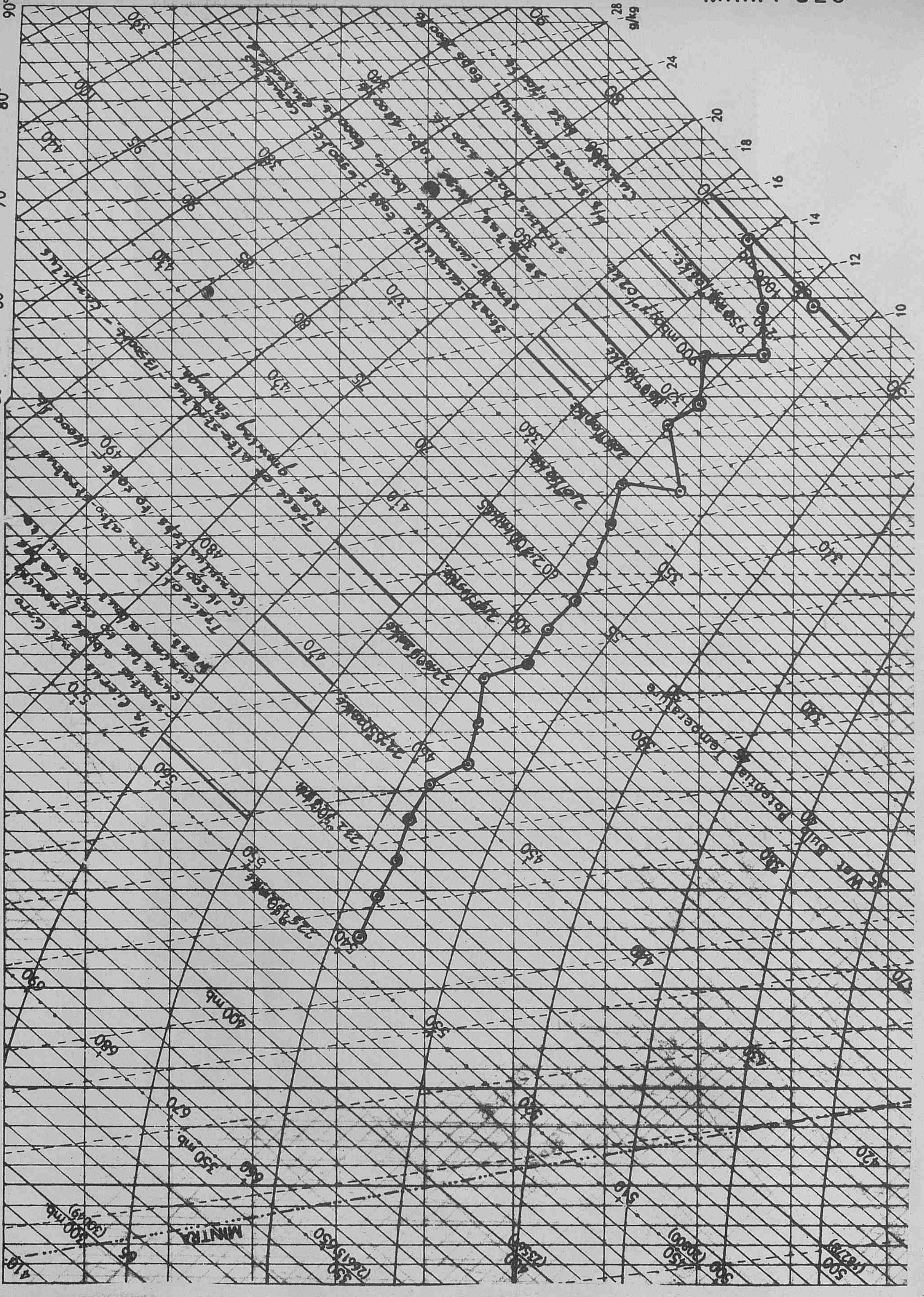


Fig. 3a   Mosquito ascent over Cranfield on 6-8-52 at about 12.00Z.

Flight 2. Figs. 4a and b.

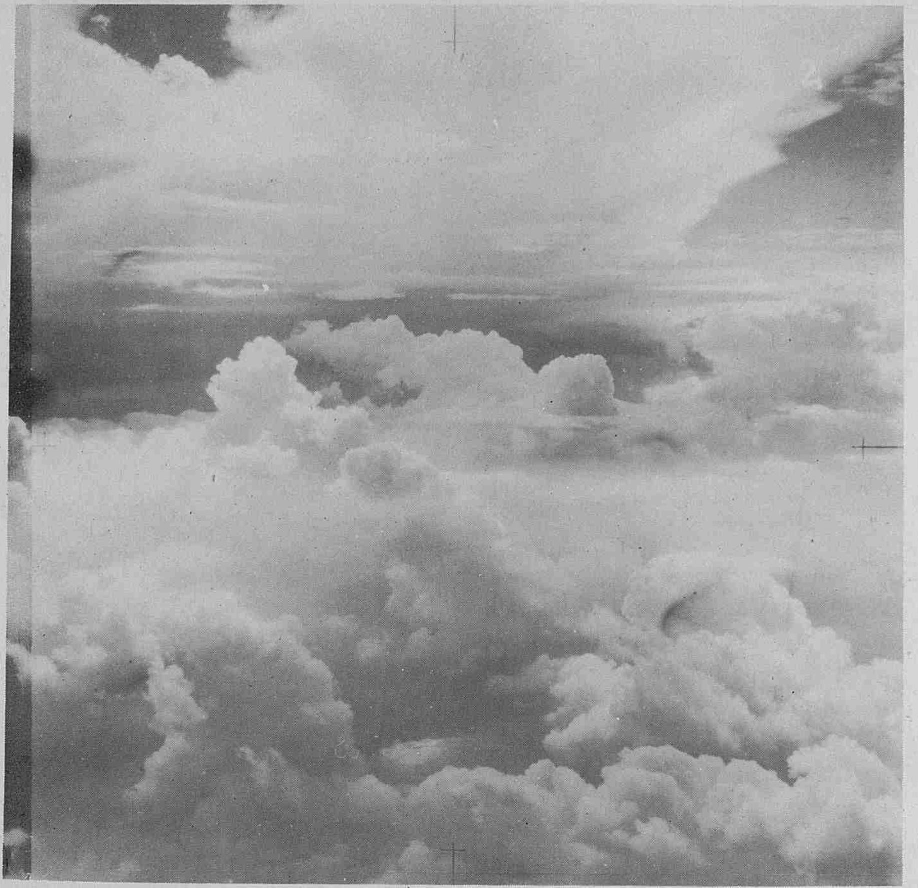


Fig. 4a. Developing cumulus from 7500 ft. overshadowed by anvil of Fig. 4b.

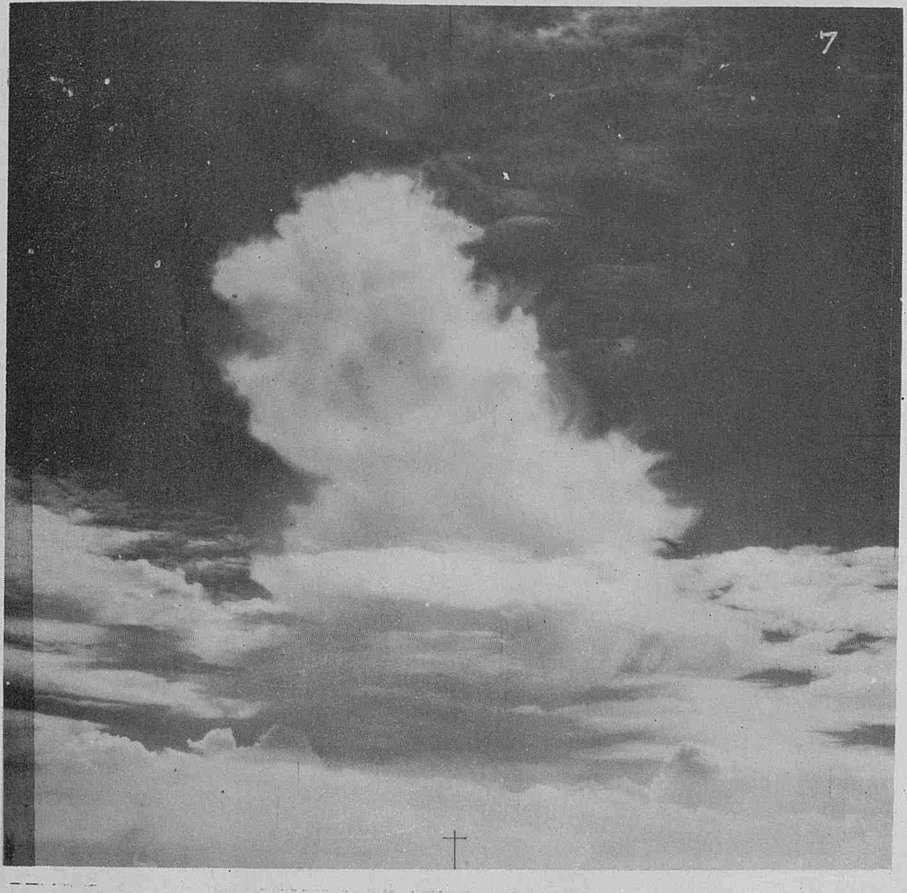


Fig. 4b. Part of distant, glaciated, cumula-nimbus top above 25000 ft.

Flight 2. Figs. 4c and 4d.

← Strato-cumulus layer at 760 mb.

Fig. 4c. Developing cumulus from 7000 ft.

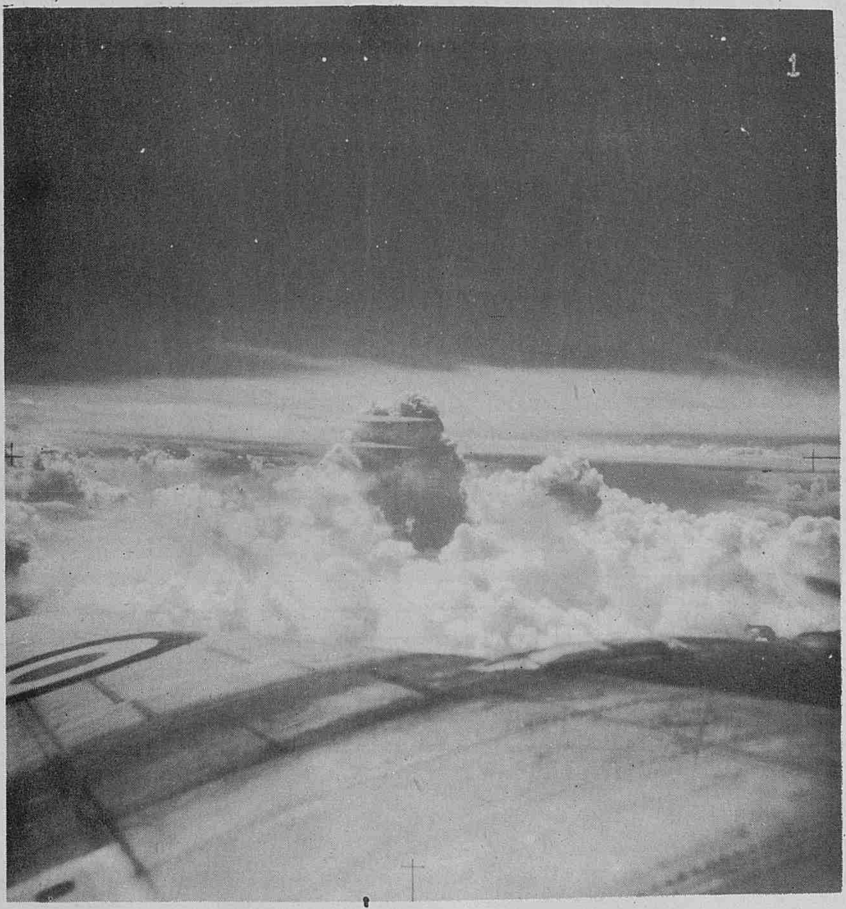


Fig. 4d. Developing Cumulus from 5000 ft.

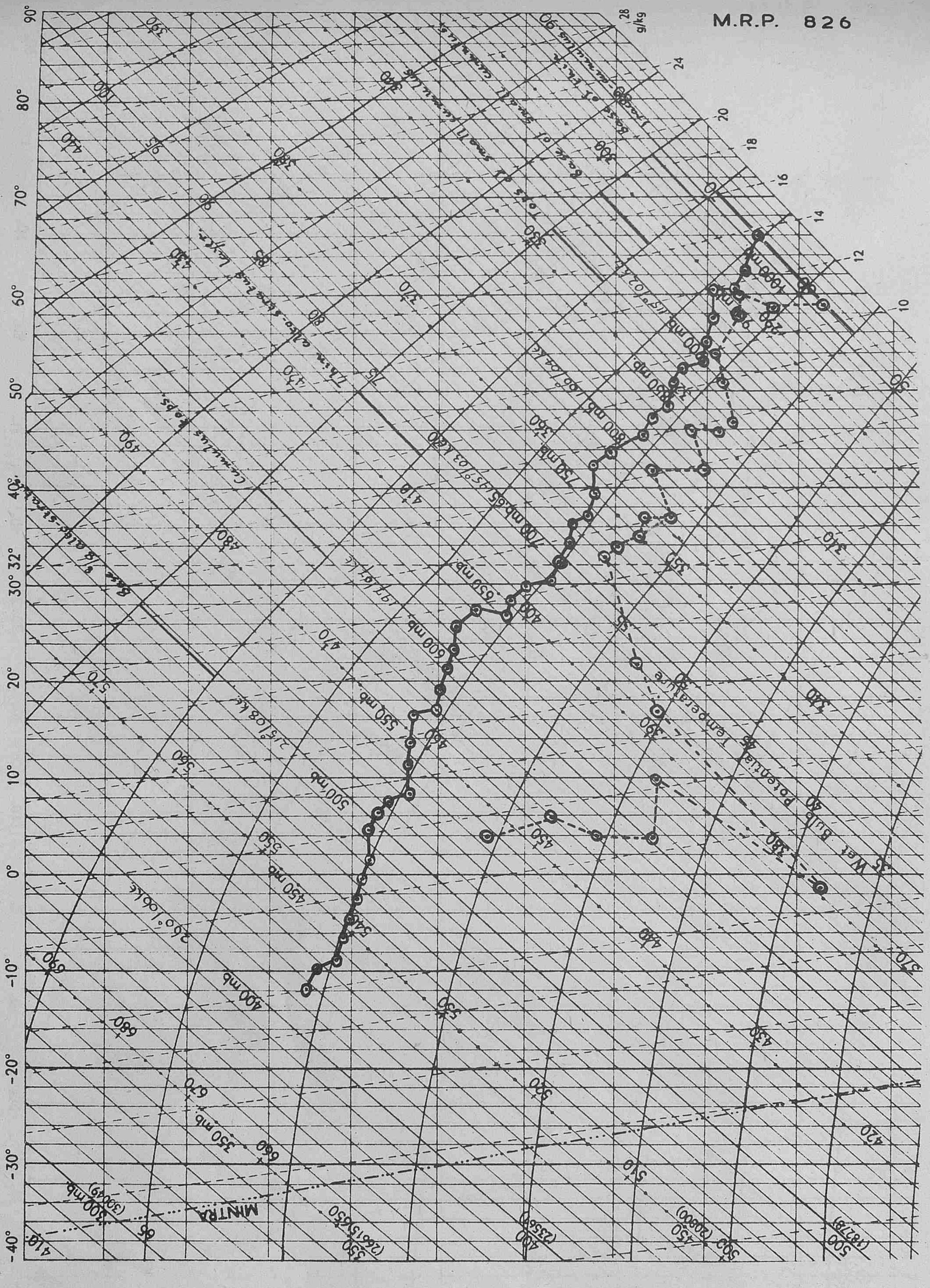


Fig. 5 Hastings ascent over Cranfield on 7.8.52 at about 11.30 Z.

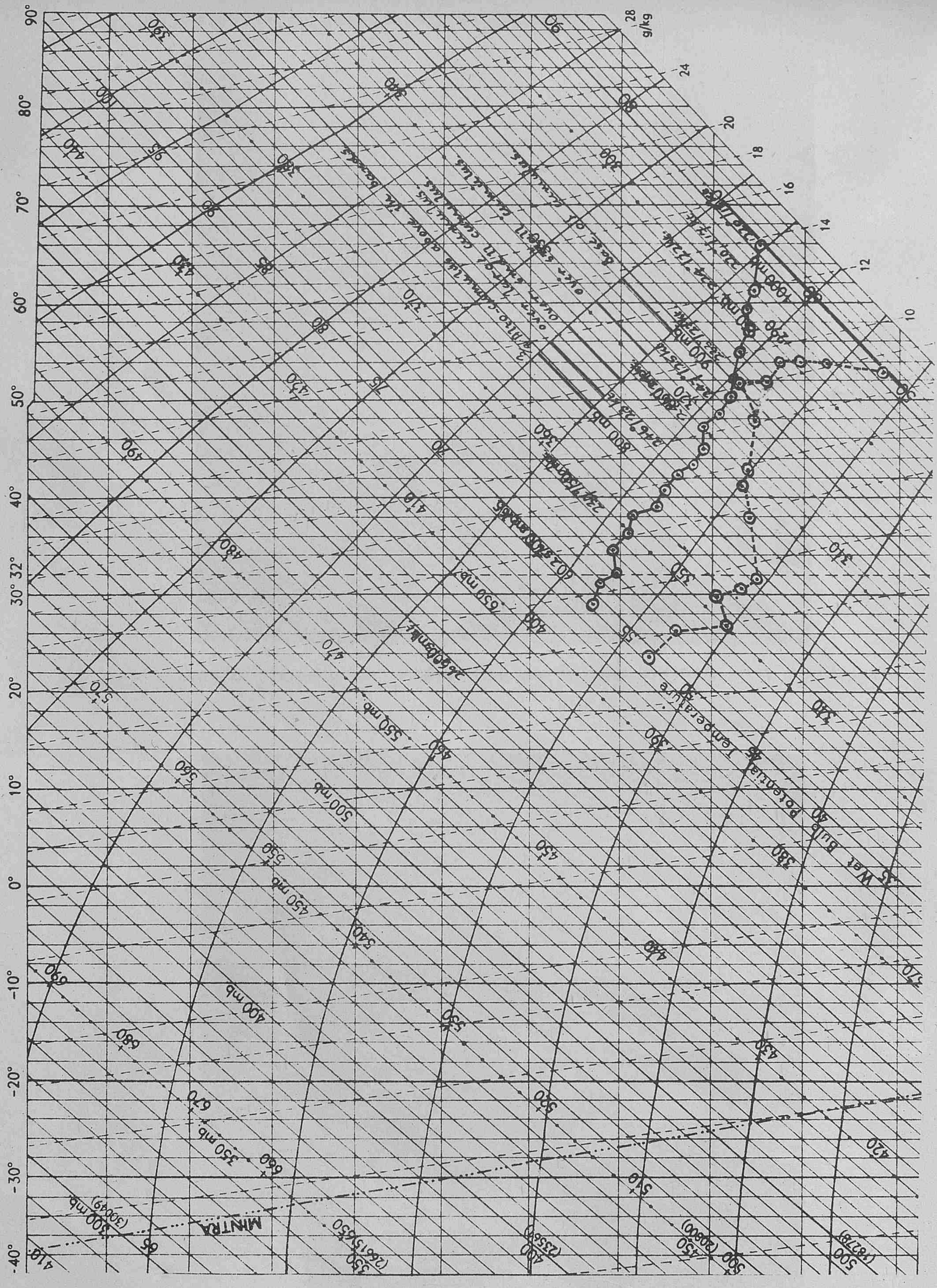


Fig. 6 Hastings ascent northwards from Farnborough on 8.8.52

at about 11.30 Z.

Flight 4. Figs. 7a and b.

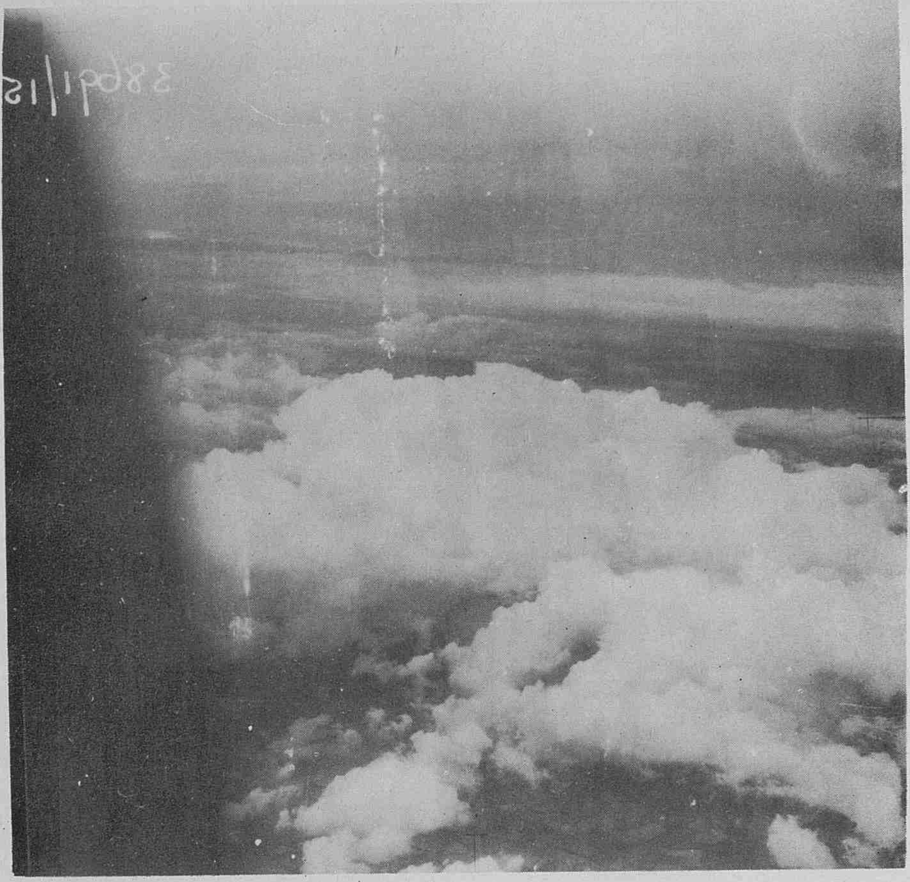


Fig. 7a. Complex investigated, viewed from 9000 ft. and probably from South

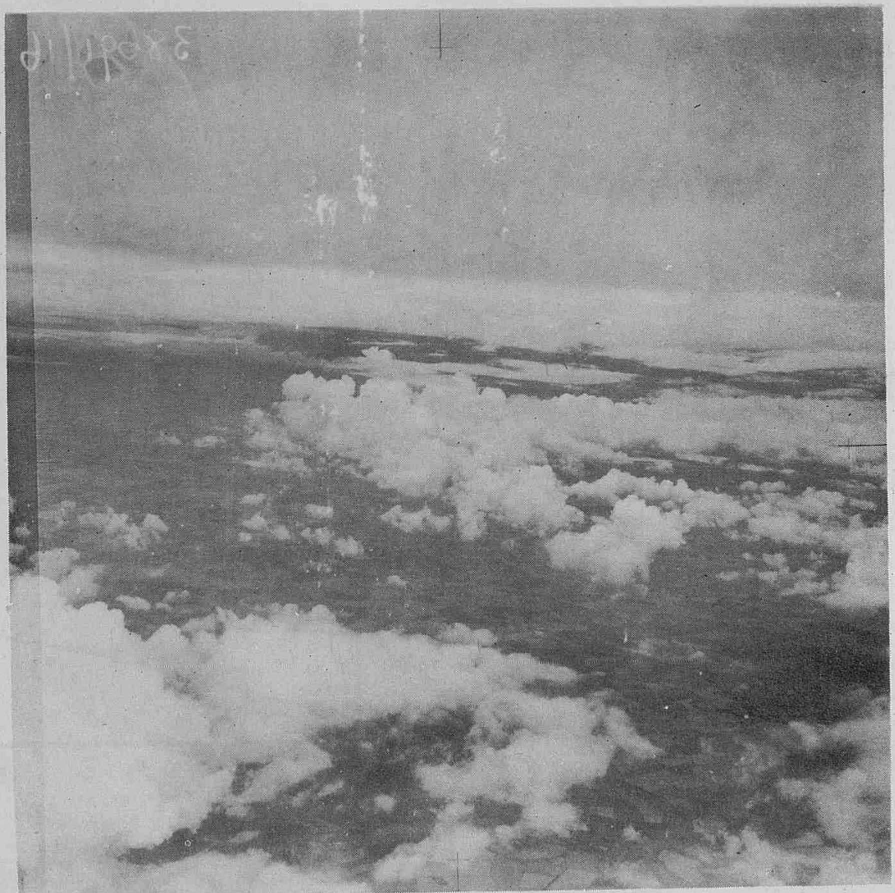
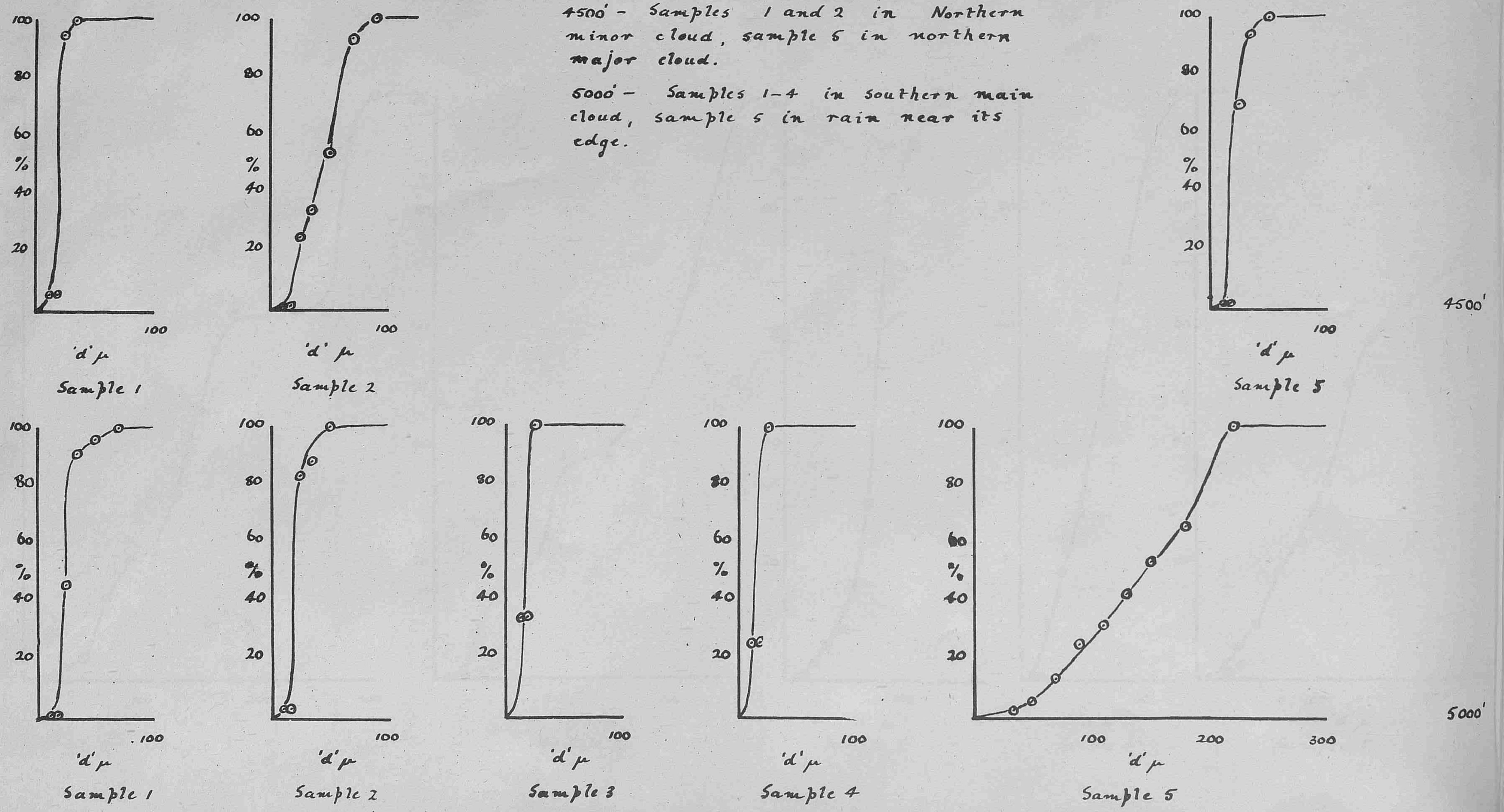


Fig. 7b. General view of cloud in area, complex of Fig. 7a on left.

Fig. 8 Droplet distribution by volume - Flight 4

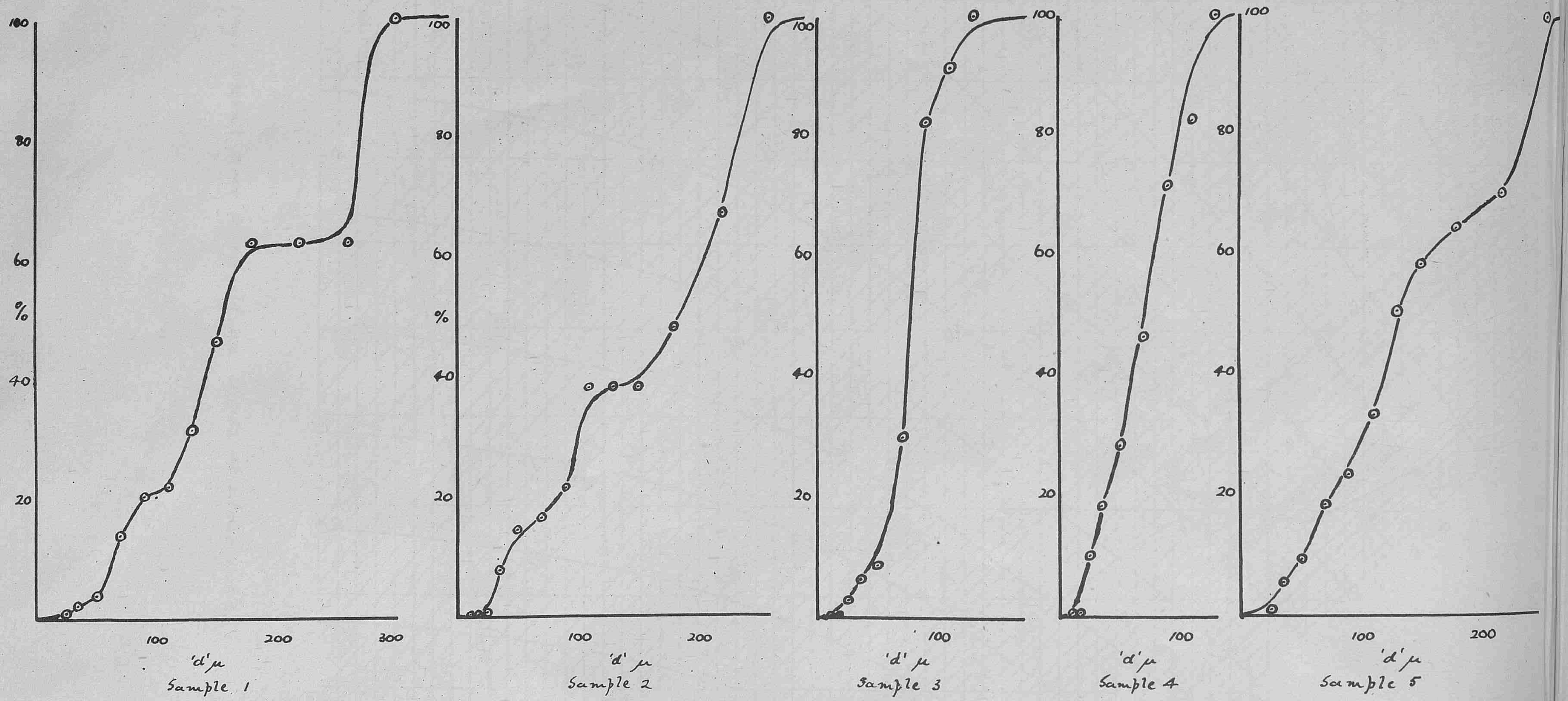


4500' - Samples 1 and 2 in Northern minor cloud, sample 6 in northern major cloud.

5000' - Samples 1-4 in southern main cloud, sample 5 in rain near its edge.

Fig. 8 (contd) Droplet distribution by volume - Flight 4

Samples in rain below and close to cloud base.



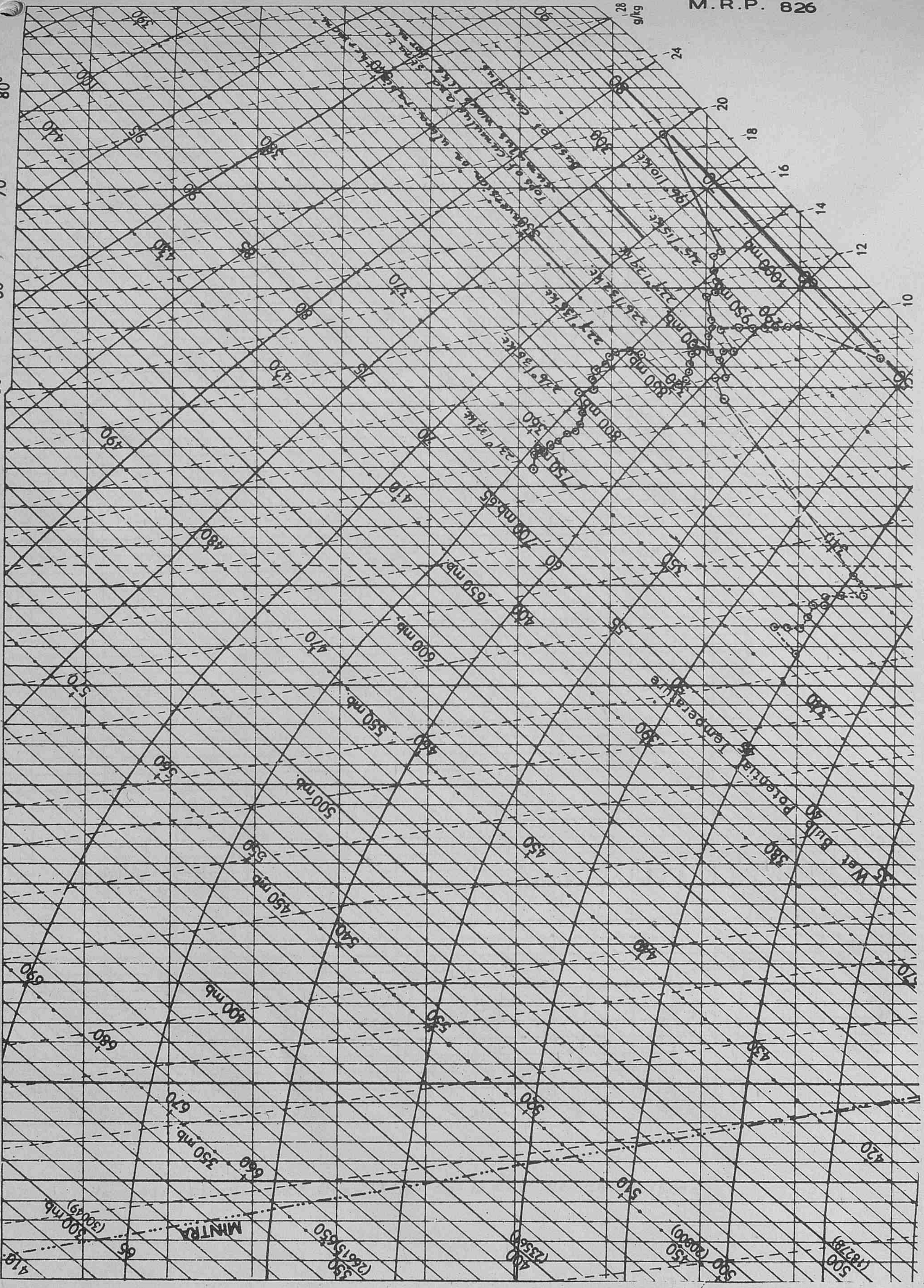


Fig. 9 Hastings ascent over Cranfield on 12.8.52 at about 1100 Z.

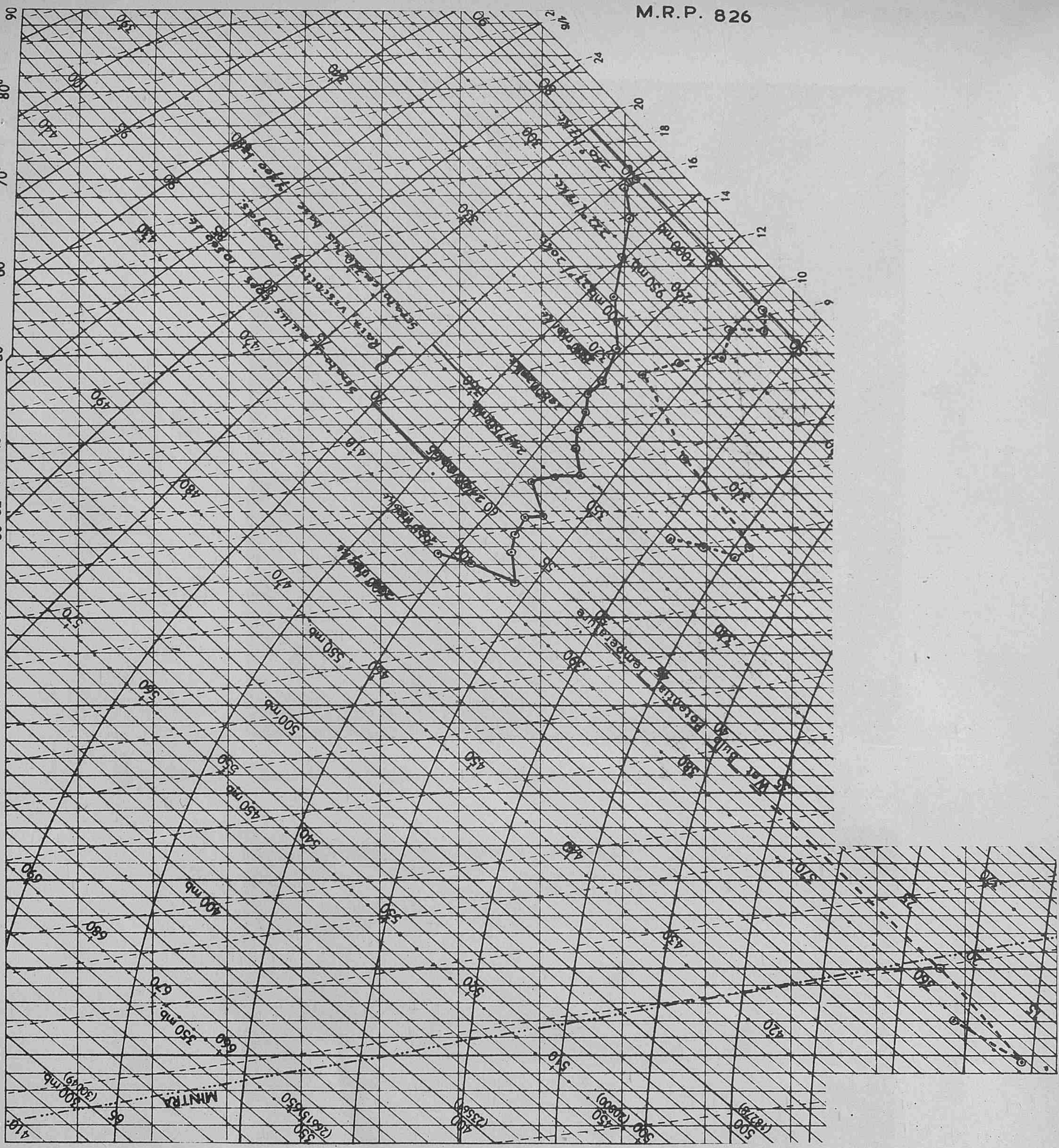


Fig. 10 Hastings ascent over Cranfield on 13-8-52 at about 1500Z.

Flight 6 – Figs. 11a and b

Fig. 11a. Cumulus penetrating strato-cumulus sheet.

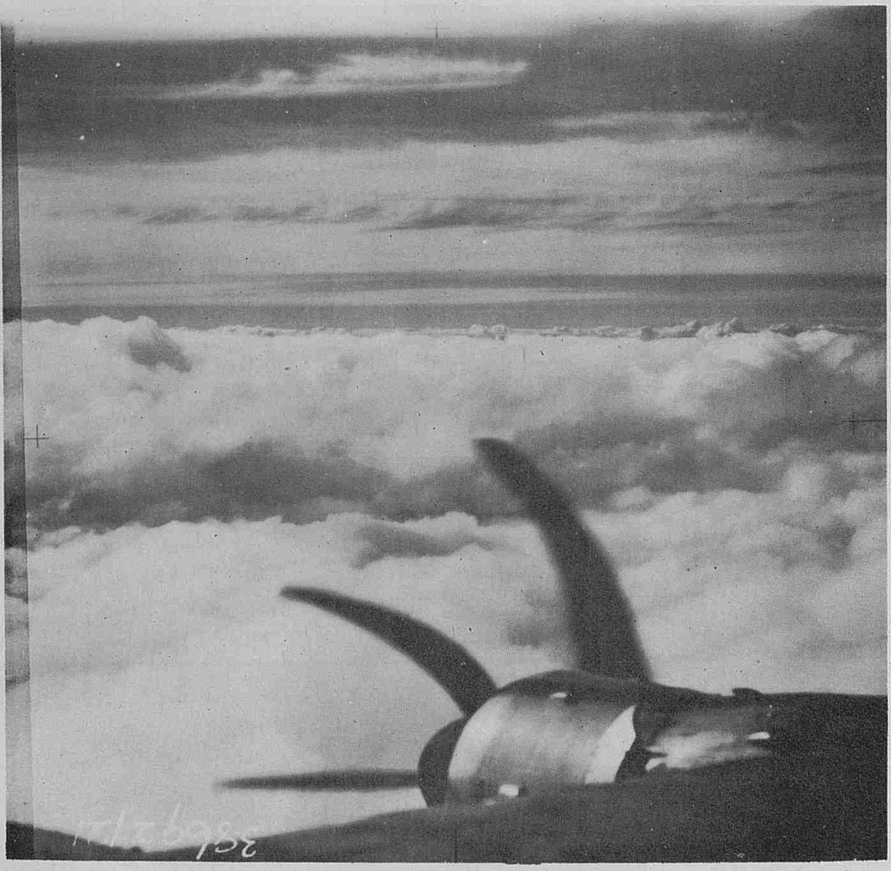
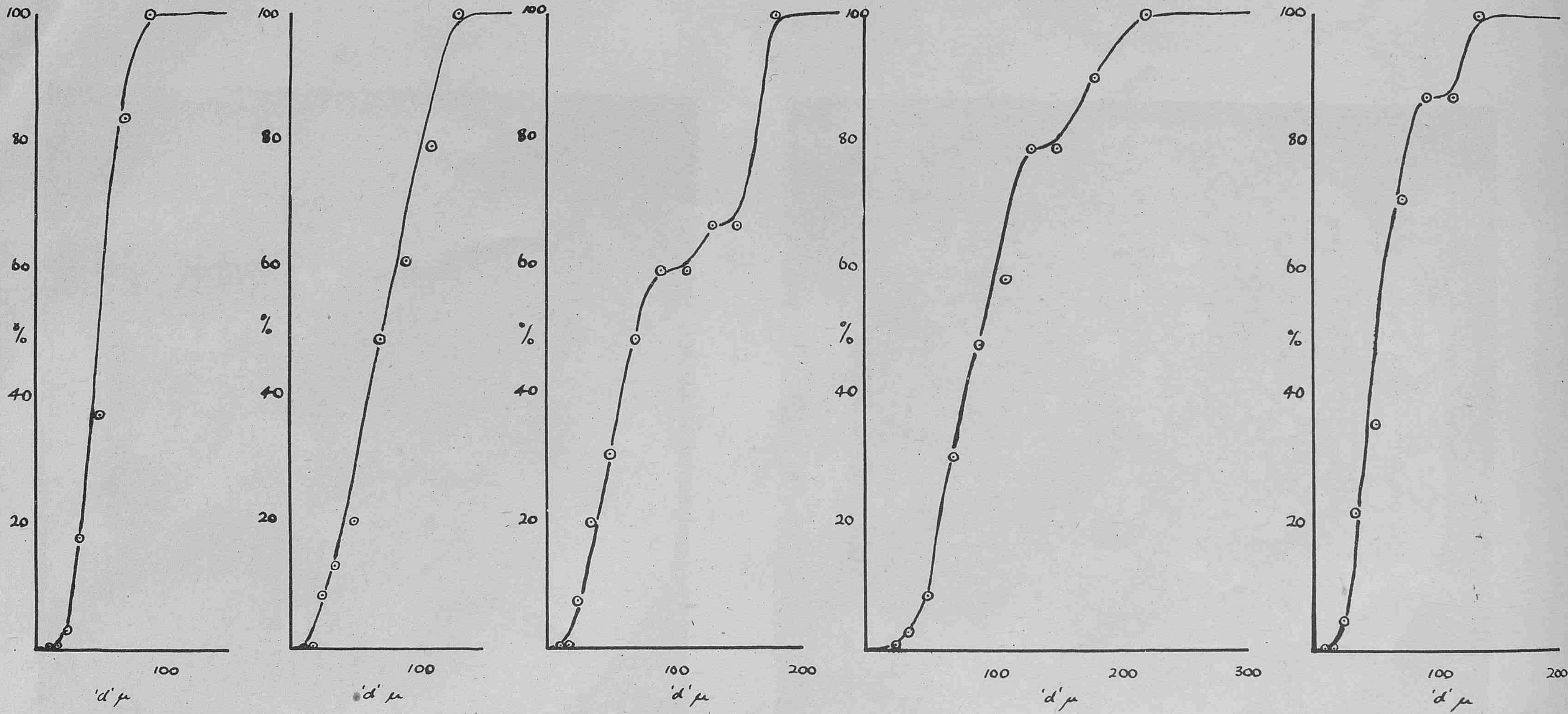


Fig. 11b General view of cloud sheet investigated.

Fig. 12 Droplet distribution by volume - Flight 6



Sample 1 Sample 2 Sample 3 Sample 4 Sample 5

Basic sample separation 200 yd.

8000'

Flight 6 - Figs. 13a and 13b.

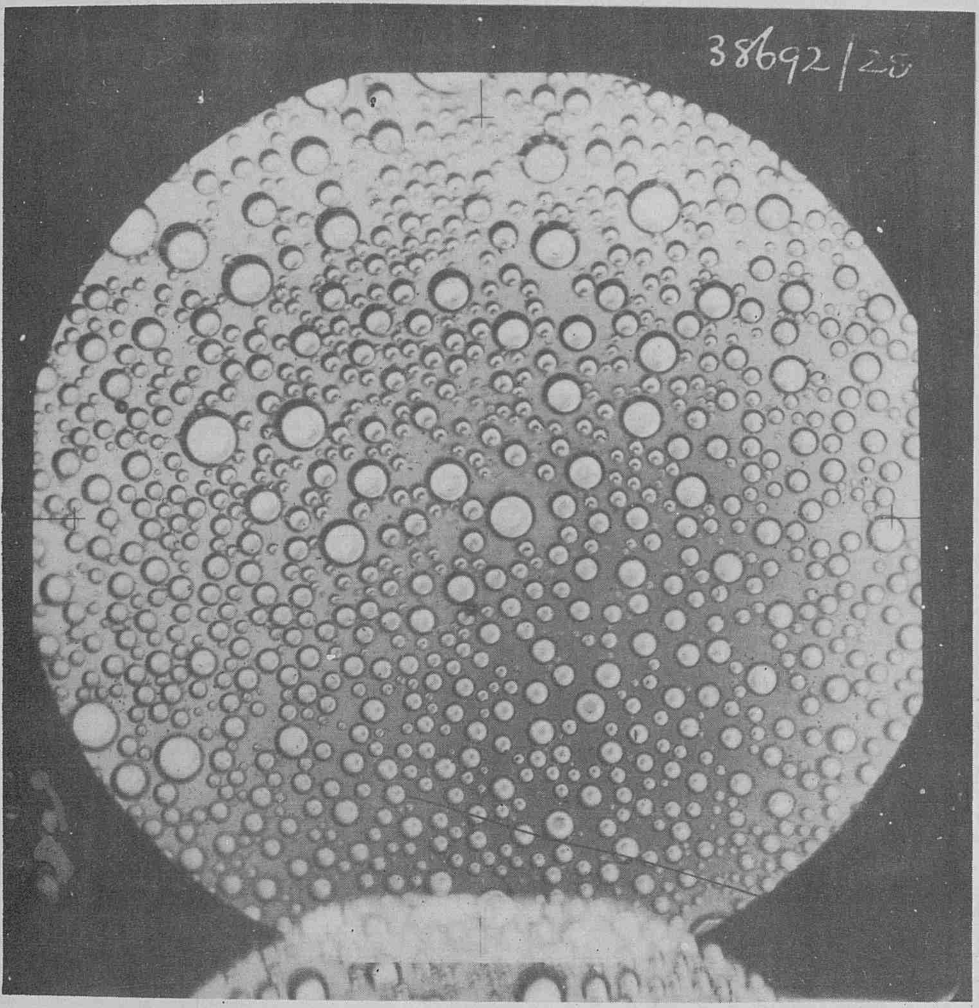


Fig. 13a.

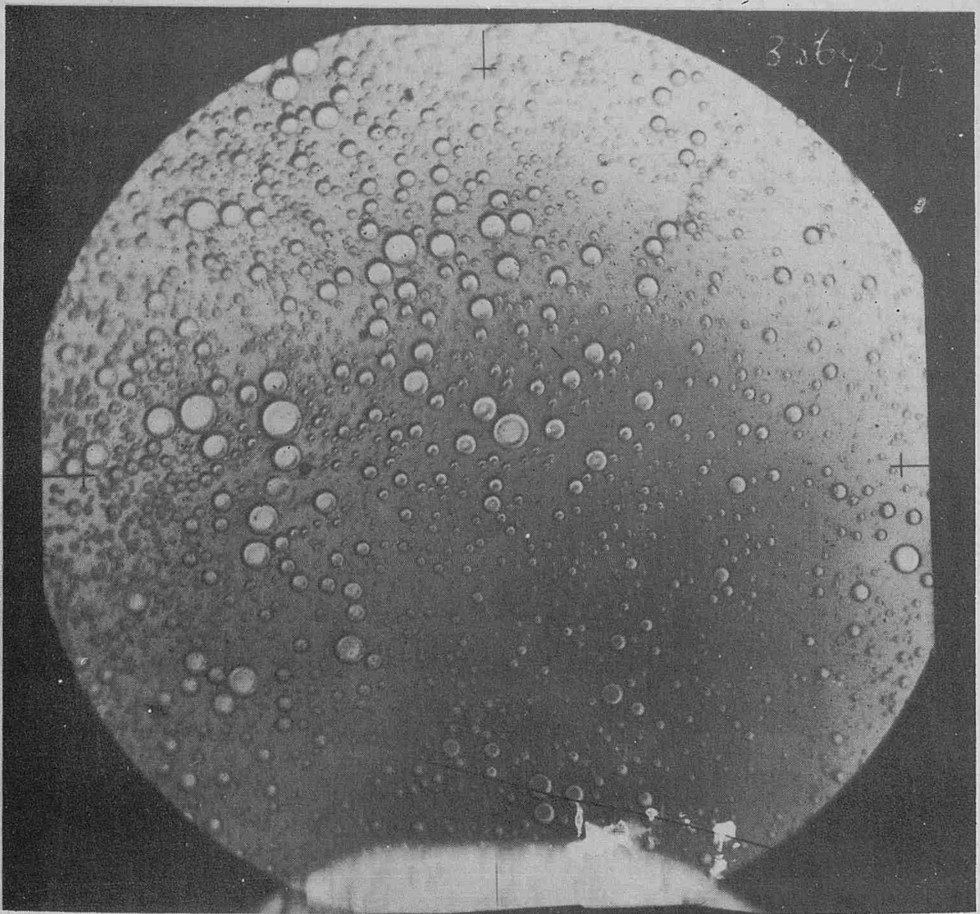


Fig. 13b. Typical examples of droplet samples

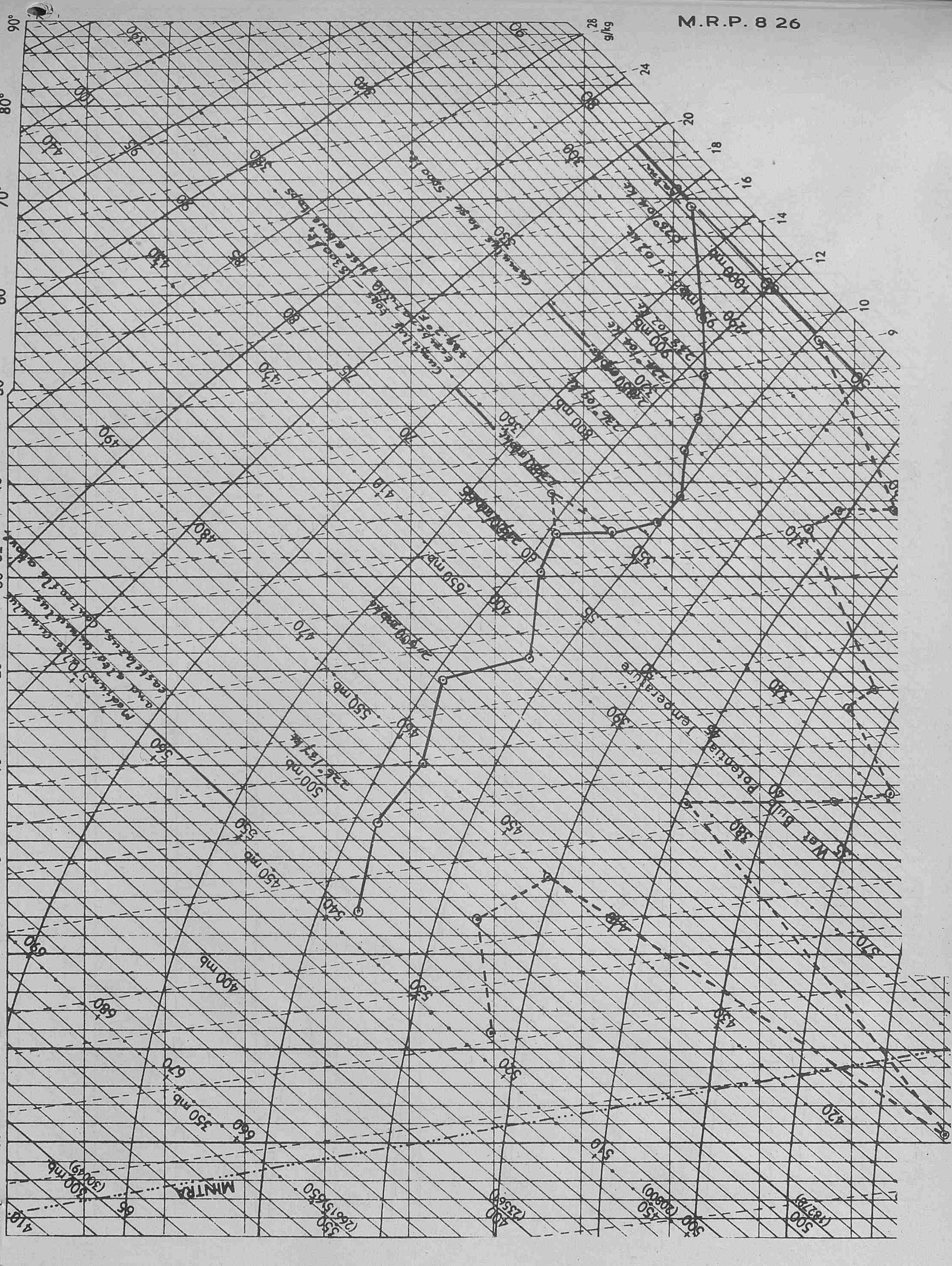


*Fig. 14   Mosquito ascent over Cranfield on 14·8·'52 at about 10·00 Z.*

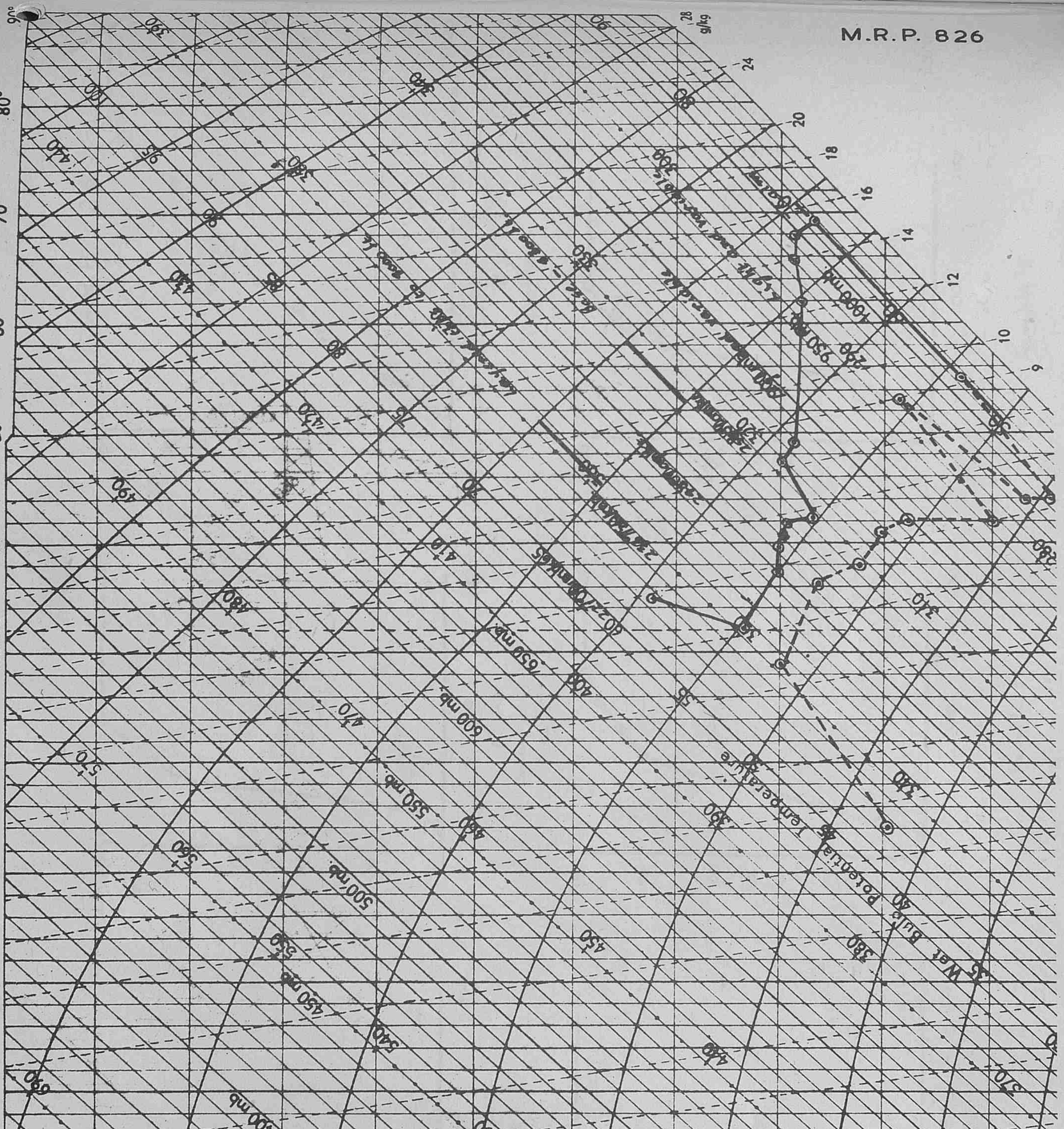
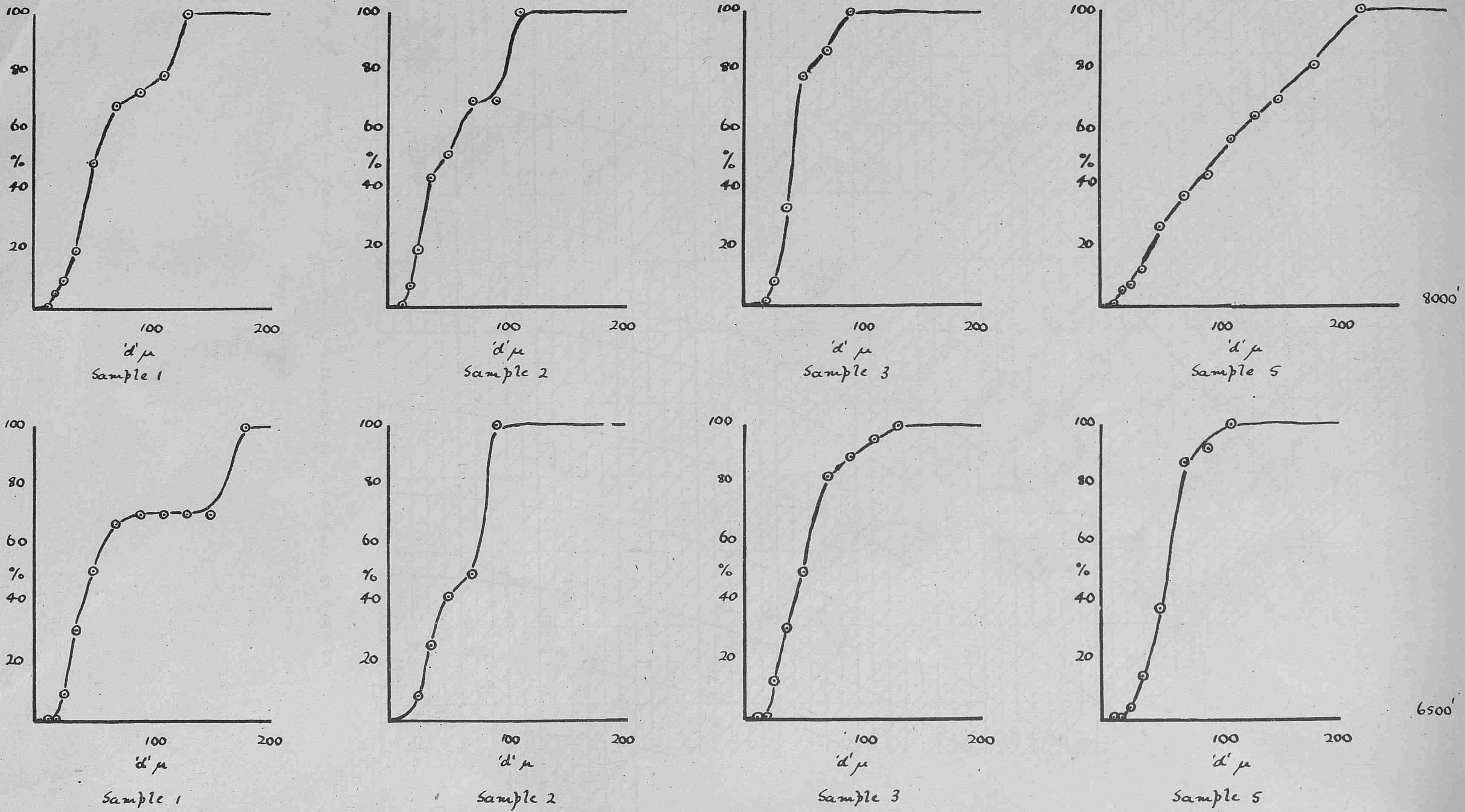


Fig. 15 Hastings ascent over Cranfield on 14.8.52 at about 1500 Z.

Fig. 16 Droplet distribution by volume - Flight 7



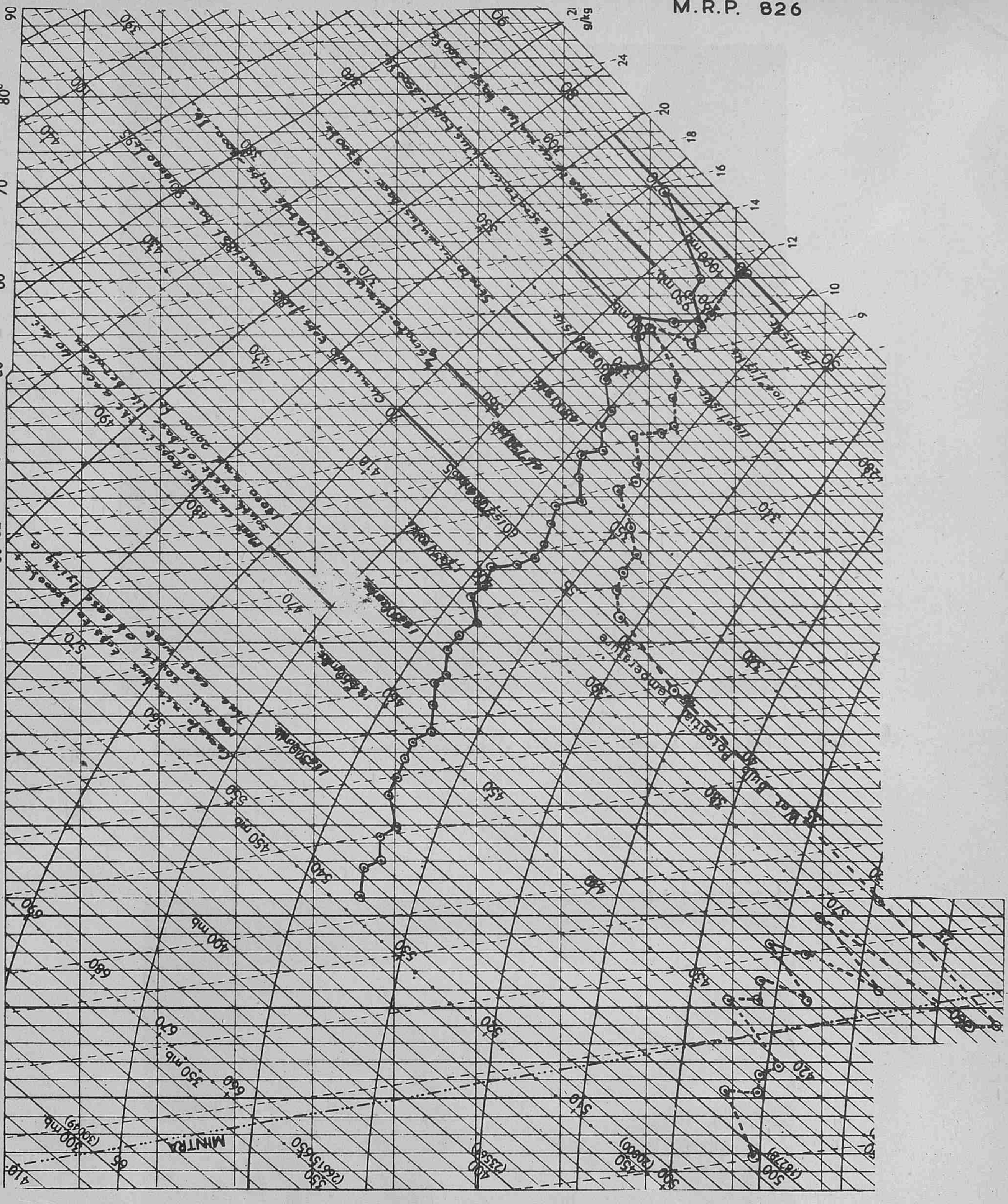


Fig. 17. Ascent in Hastings heading south-westerly from base on 15.8.'52 about 1400 Z.

Flight 8 - Figs. 18a and b.

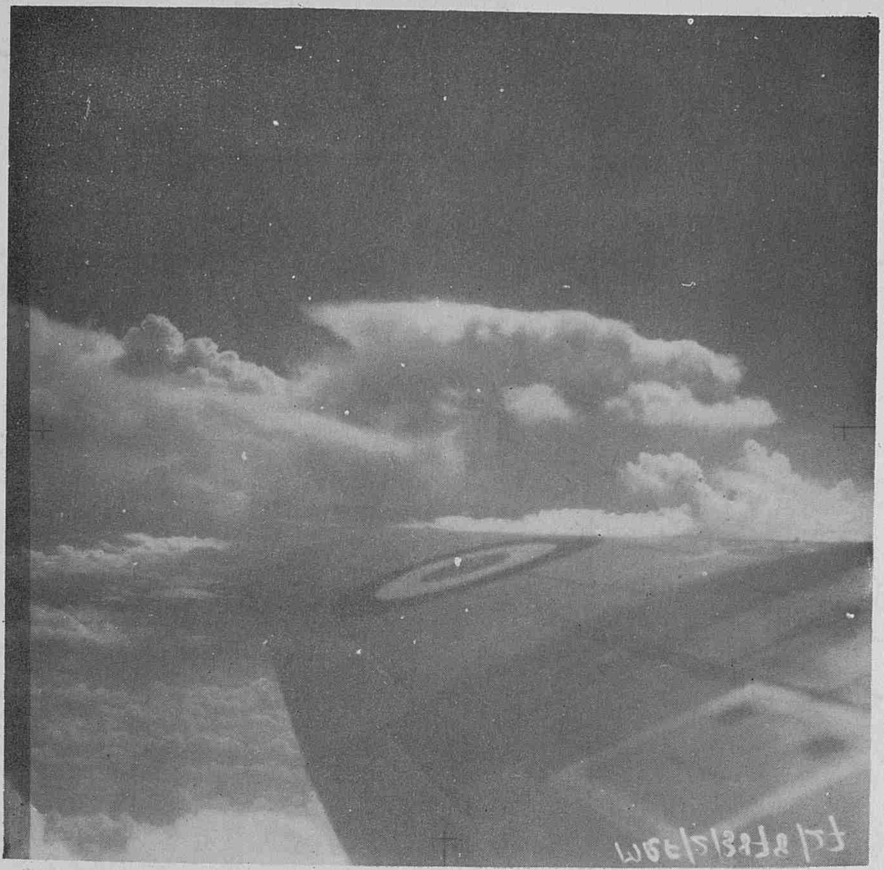


Fig. 18a. Cumulo-nimbus cloud originally chosen, anvil at about 30000 ft.

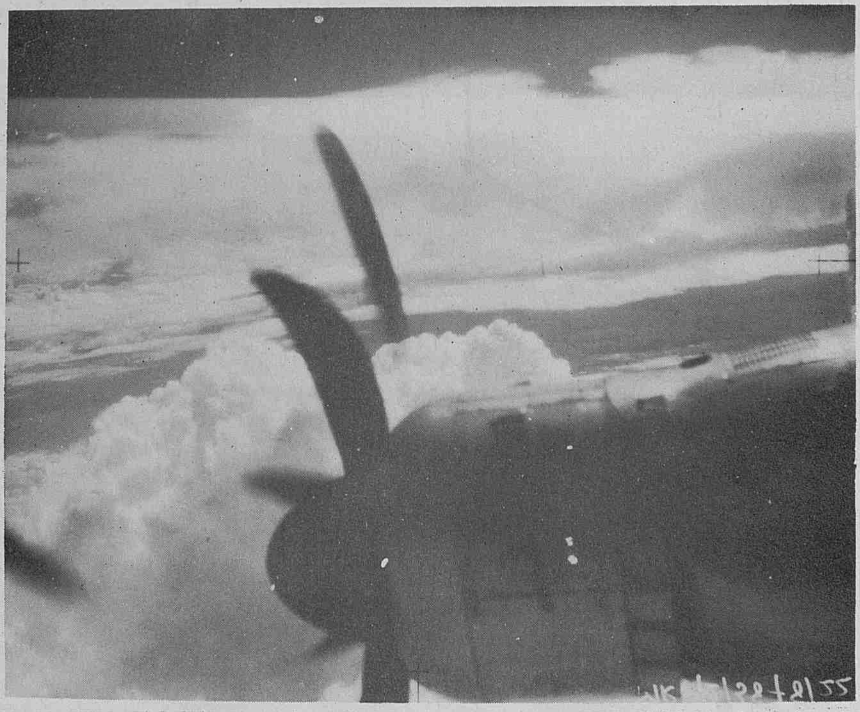
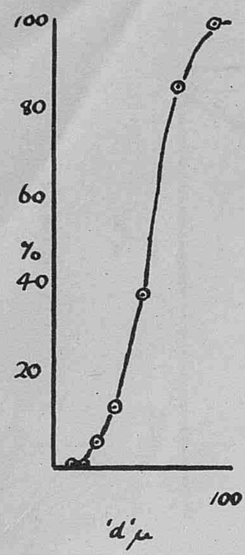


Fig. 18b. Cloud investigated, top at about 18000 ft.

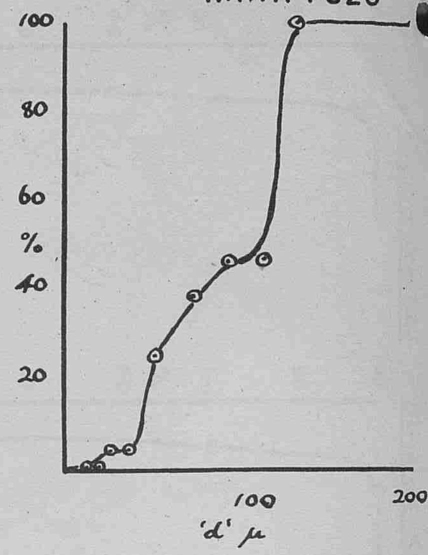
Fig. 19 Droplet distribution by volume - Flight 8

Basic sample seperation about 200 yd.

17,000'

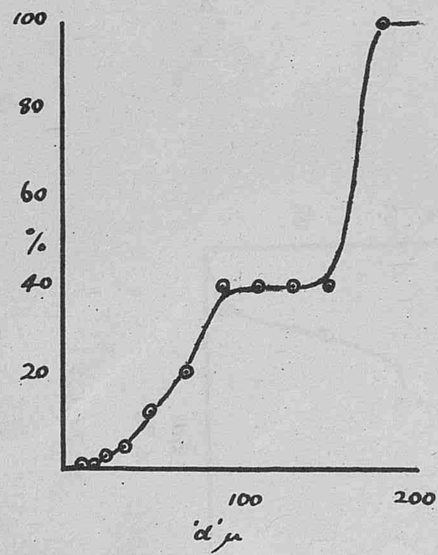


Sample 1

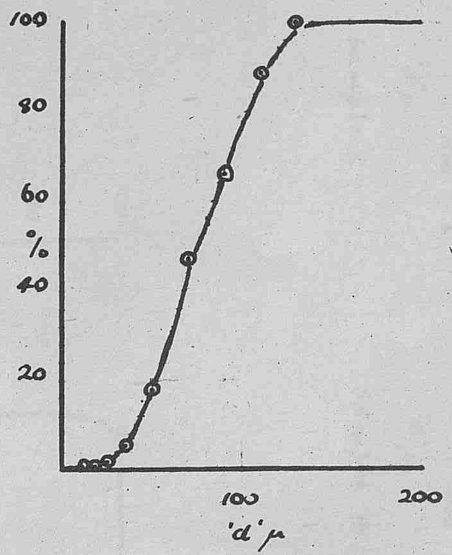


Sample 4

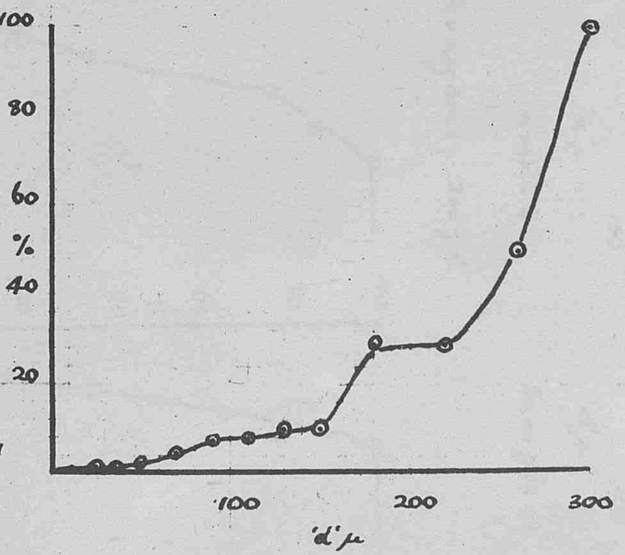
15,000'



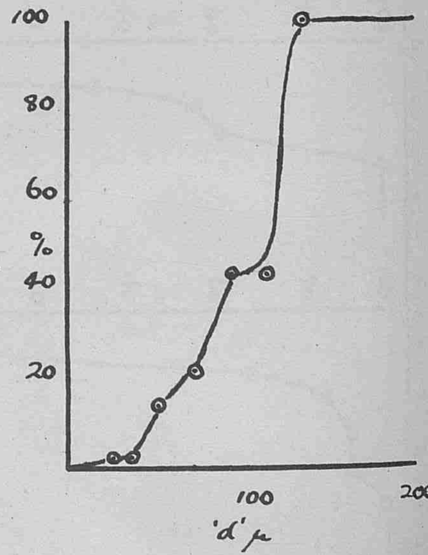
Sample 1



Sample 2

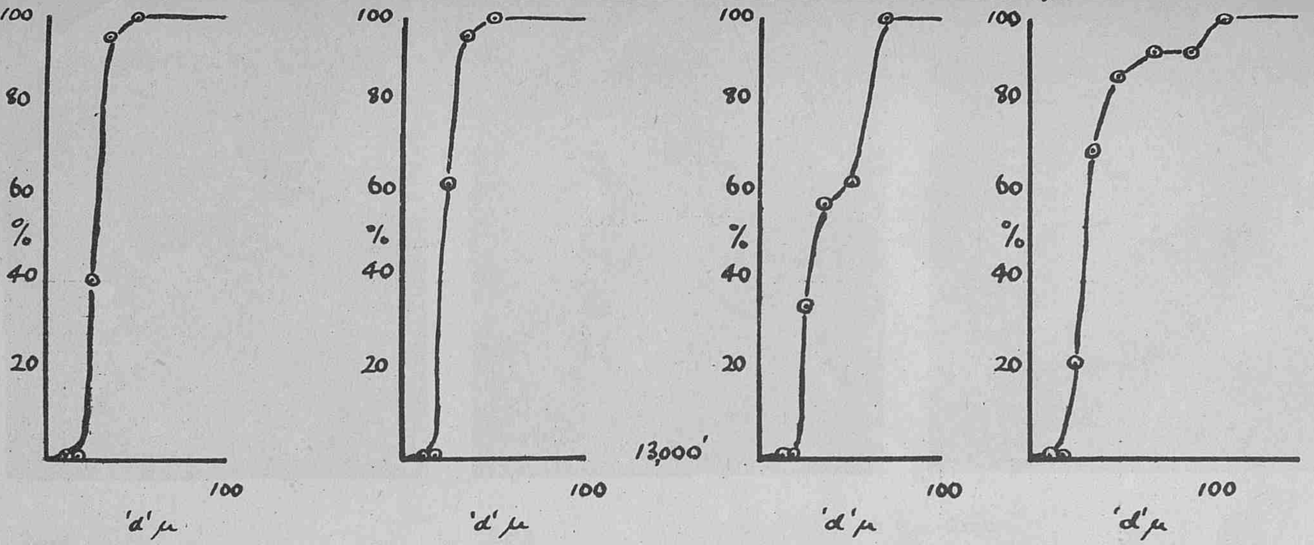


Sample 3



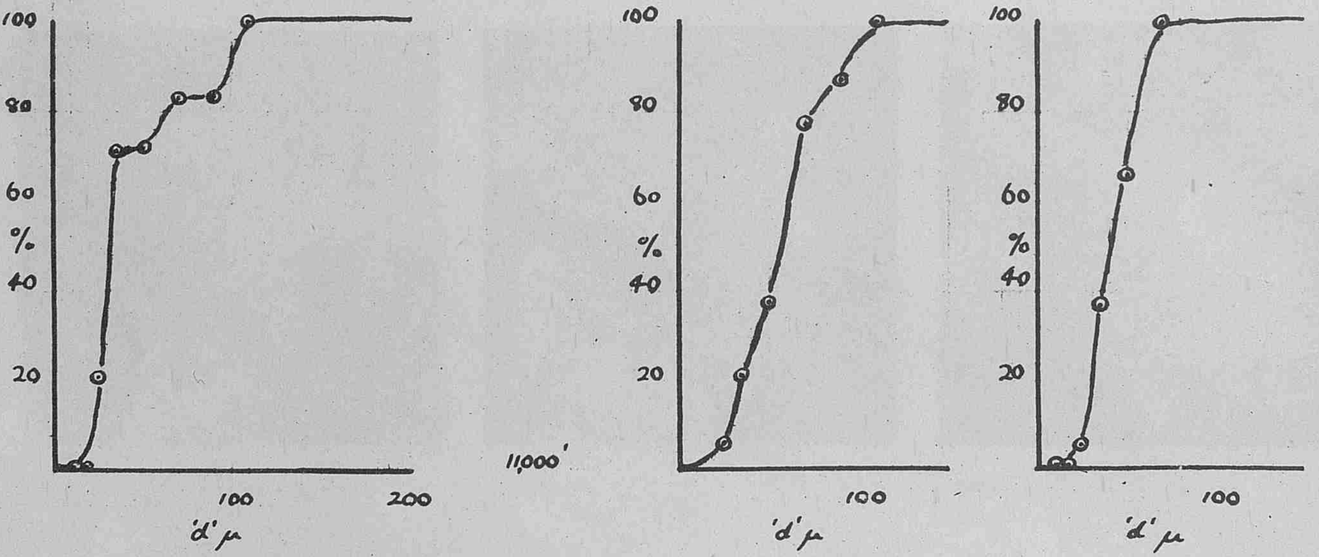
Sample 4

Fig. 19 (contd.) Droplet distribution by volume - Flight 8

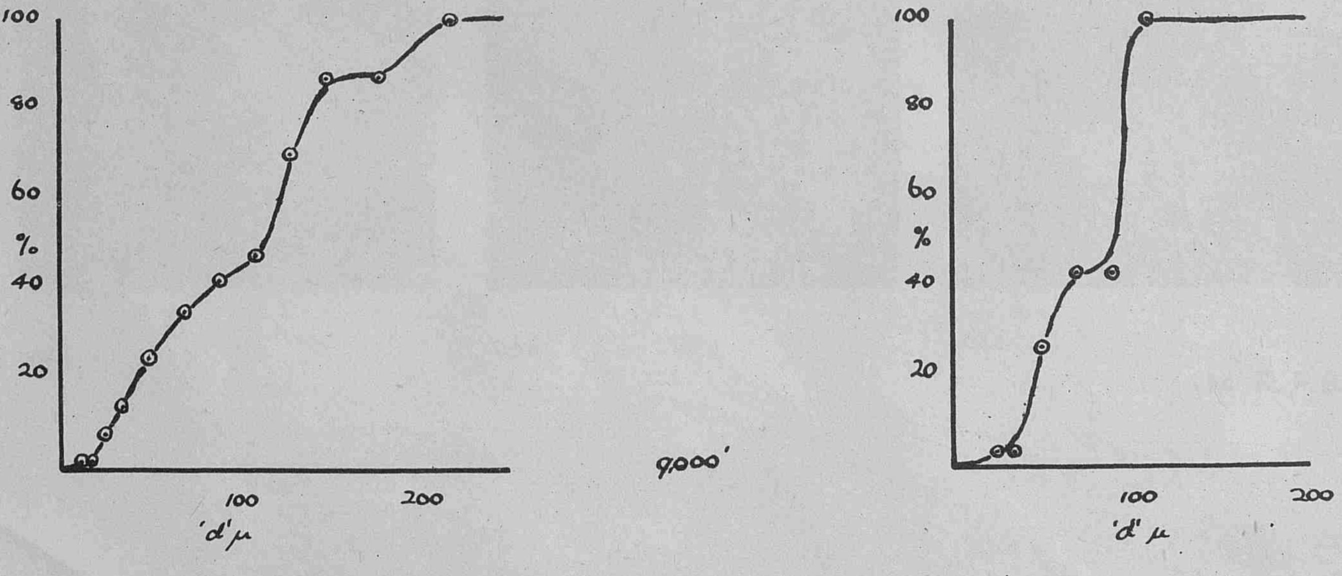


Sample 1 Sample 2 Sample 3 Sample 4

Basic sample seperation throughout 200 yd.

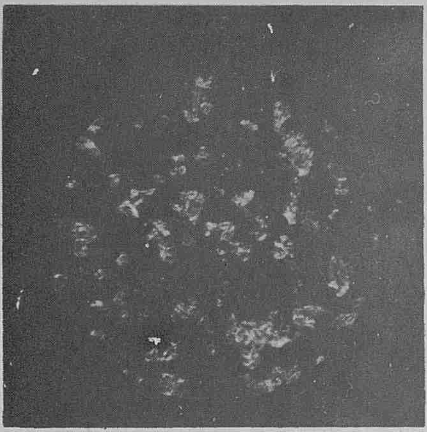


Sample 1 Sample 3 Sample 4

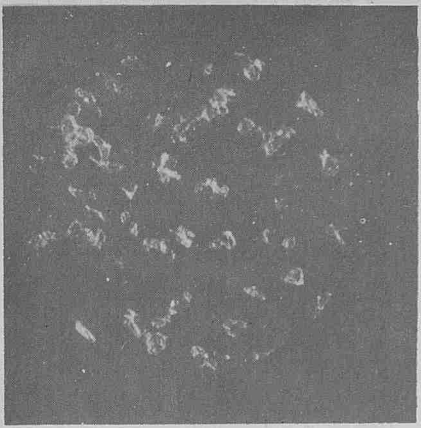


Sample 1 Sample 4

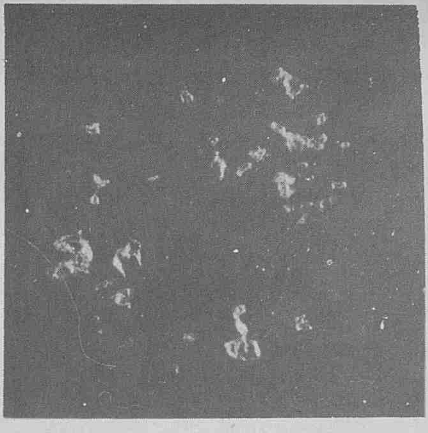
Figs. 20a–i. Typical Crystal samples



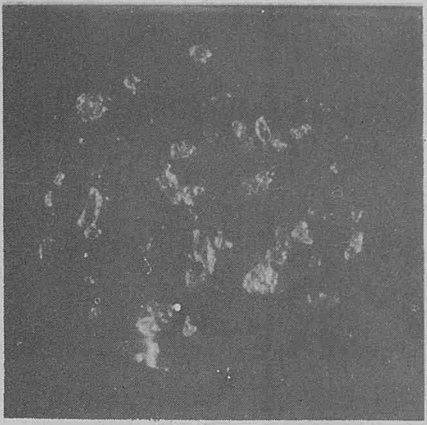
20a.



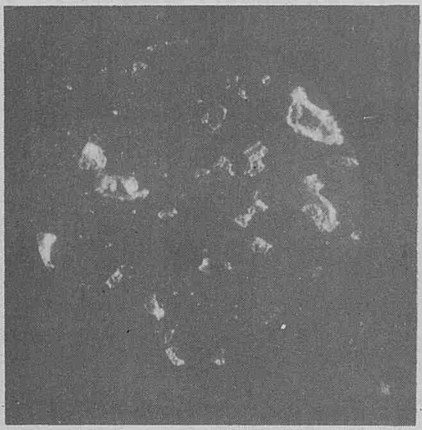
20b.



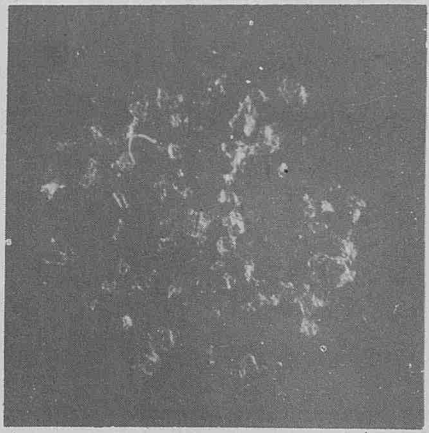
20c.



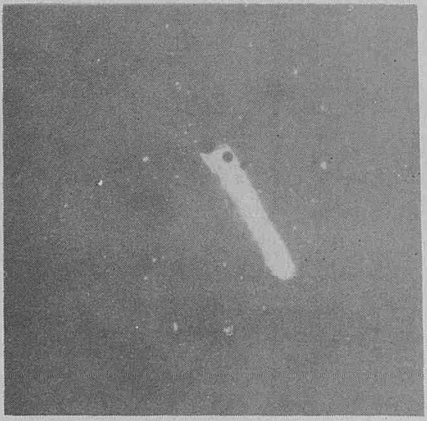
20d.



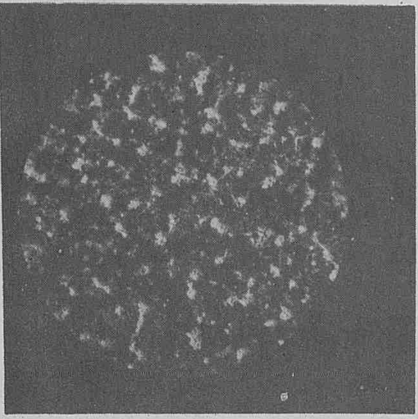
20e



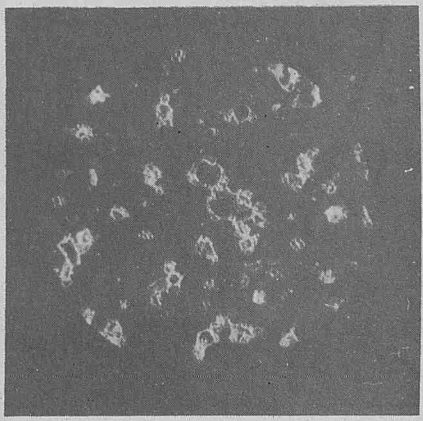
20f



20g.



20h.



20i.

At the time of issue this report has not been [illegible]. Any [illegible] desiring to [illegible]
[illegible]
copy of [illegible] is available in the library of
The Meteorological Office.)

M.R.P. 827
[illegible]

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

## Investigation of high altitude clear air turbulence near jet streams: special flights by R.A.F. and [illegible] aircraft

by

[illegible] (Meteorological Office)

**Summary**

[illegible]

[illegible] high altitude clear air turbulence [illegible] strong shear of wind in the vertical [illegible] horizontal or isentropic surfaces.

A recent analysis of reports of severe turbulence encountered by aircraft over the United States (1) showed that nearly 70 per cent of the incidents occurred in association with well defined jet streams. The others were associated with upper lows, upper troughs, discontinuous tropopause surfaces or strong winds not associated with jet streams but a small number had no obvious association with any configuration in the upper air pattern. [illegible] of those associated with jet streams were on the low pressure side of the stream.

2. Special flights by R.A.F. aircraft. In order to confirm or correct

At the time of issue this report has not
been published. Any author desiring to
quote this paper should refer to it as "A
paper of the Meteorological Research
Committee (London),
M.R.P. No. 827 by Mr. D.C.E. Jones
A copy is available in the Library of
the Meteorological Office."

M.R.P. 827
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24th July, 1953

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# AIR MINISTRY

# METEOROLOGICAL RESEARCH COMMITTEE

## Investigation of high altitude clear air turbulence near jet streams: special flights by R.A.F. and R.A.E. aircraft

by

D.C.E. Jones (Meteorological Office)

### Summary

Special investigational flights were made by Royal Air Force and R.A.E. jet aircraft to explore and report the variations in the occurrence and severity of clear air turbulence in different parts of jet streams over the British Isles. Apart from a few exceptions it was found that moderate or severe turbulence occurred mainly on the low pressure side of the axis and to a lesser extent on the high pressure side above the axis; very few incidents were reported on the high pressure side below the axis. It was tentatively deduced that there was an approximately equal probability that the worst bumpiness encountered near a jet stream was either severe, moderate or perceptible, and that a small proportion of jet streams were not turbulent at all. Generally severe turbulence occurred when the vector difference between the wind at 300 and 500 mb. levels at the jet exceeded 35 knots and the horizontal shear on the low pressure side at the level of the axis of the jet exceeded 0.2 hr.$^{-1}$

1. Introduction. There is a considerable amount of statistical information now available concerning the frequency, vertical and horizontal extent, strength and levels of occurrence of turbulence over the British Isles and Western Europe at heights between 15,000 and 40,000 ft. (1, 2). High altitude clear air turbulence has been observed in almost all weather situations, though it appears that there are two factors, one or other or both of which are usually associated with its occurrence, namely, strong shear of wind in the vertical and strong shear of wind in the horizontal or isentropic surfaces.

A recent analysis of reports of severe turbulence encountered by aircraft over the United Kingdom (3) showed that nearly 70 per cent of the incidents occurred in association with well defined jet streams. The others were associated with upper lows, upper troughs, discontinuous tropopause surfaces or strong winds not associated with jet streams but a small number had no obvious association with any configuration in the upper air pattern. 75% of those associated with jet streams were on the low pressure side of the stream.

2. Special flights by R.A.F. aircraft. In order to confirm or correct certain aspects of the above results, and to study further the occurrence of turbulence near jet streams it was agreed that at certain airfields the Royal Air Force would make investigational flights in areas specified by the Meteorological Office. The investigations were made by the dispatch of a special aircraft, or by diverting an aircraft already airborne and, at times, it was made in conjunction with the normal daily flying programme of the station.

But the amount of flying effort available was subject to the proviso that operational and training commitments took first priority, which was often most unfortunate from the point of view of the investigation because many opportunities of flying and investigating turbulence in interesting weather situations were unavoidably missed.

This investigation was given the code name "Upper Bumps" and was started in March 1952. Reports obtained during the period up to February 1953 have been studied and are dealt with in this paper.

The main interest was the jet stream with which turbulence at high levels is now known to be mostly associated. The Meteorological Officer at each of the airfields nominated to take part in the investigation informed the Wing Commander Flying when the conditions appeared favourable for the occurrence of turbulence, giving as much advance notification as possible. A request for an investigational flight was then made and the approximate area or route, and the height of flight specified.

The pilot undertaking the flight was asked to note the occurrence of turbulence throughout the flight, observations from height above 20,000 feet being the main interest. Nil reports were regarded as being just as important as positive reports. Details were required of the severity of the turbulence, that is, whether it could be classed as, perceptible, moderate or severe.

The aircraft used were not fitted with any special instruments and therefore the reports made by the pilots could only be qualitative assessments. In order to achieve as much consistency as possible with regard to the severity of the turbulence encountered, all pilots were asked to pay careful attention to a scale which has been described in reference (4).

Pilots carrying out these investigational flights were encouraged to report special or unusual features of the turbulence and to state if any action was taken either by climbing or descending or altering course or speed, to fly out of turbulent areas. The boundaries of the turbulent zones were discovered and noted wherever possible. No attempt was made to carry out a systematic investigation of the variations along the jet streams; only the cross sections of the jets were explored.

At the conclusion of each investigational flight the pilot was "debriefed" by the Meteorological Officer on the station and a special form was completed. An example of a completed form is given in Appendix I.

In all 23 jet streams occurring over the British Isles were investigated during the first year that Operation "Upper Bumps" was carried out. The number of flights made either across or in particular regions of jet streams varied from one to twelve per jet, the average being about three. It was not possible for the whole cross section of a jet stream to be thoroughly investigated in each case.

3. Analysis of results. The track of each flight was plotted on a graph showing the position of the flight relative to the jet stream. The method adopted was similar to that in reference (3), except that mean isopleths of wind speed have not been inserted. Vertical units are the difference between the pressure at the axis of the jet and the pressure on the I.C.A.N. scale corresponding to the height of the aircraft. The horizontal unit is the distance from the jet centre to the point at the same level on the low pressure side of the jet at which the wind speed decreased to half its maximum value. Figs. 1 and 2 show diagrammatically all flights carried out during the periods March to August 1952 and September 1952 to February 1953 respectively.

An attempt was made to classify each jet according to vertical and horizontal shear and to ascertain if there was any correlation between these two parameters and the intensity of the turbulence experienced. The vertical shear was taken as the shear between the 300 and 500 mb. levels at the axis of the jet, these levels being chosen because they are standard levels used for constructing contour charts. This method of evaluating the vertical shear suffers from a serious disadvantage in that it does not give a representative value when the axis of the jet lies about halfway between the two levels but such cases were quite few.

The horizontal shear was estimated on the low pressure side of the jet and at the same level as the axis of the jet. The value obtained refers to the average shear over the distance over which the wind falls to half its maximum value at the axis of the jet. In practically all cases the radius of curvature of the stream lines was so large that no adjustment was considered necessary for it.

In order to ascertain whether a relationship existed between the shear (both vertical and horizontal) and the severity of the bumpiness Figs. 3-7 were constructed. Fig. 3 shows diagrammatically the most severe intensity of turbulence encountered near each of the jet streams investigated plotted for the appropriate values of the horizontal and vertical shear. Similarly Figs. 4-7 show the most severe turbulence experienced in the four quadrants of the jet stream.

The various reports of turbulence experienced during the flights were also examined in relation to the position of occurrence along the jet stream, that is, whether they occurred near the entrance, middle or exit. The result is given in the table below:-

Table I

| | Nil | Perceptible | Moderate | Severe |
|---|---|---|---|---|
| Entrance | 0 | 5 | 0 | 0 |
| Middle | 24 | 24 | 21 | 11 |
| Exit | 0 | 10 | 12 | 2 |

It would appear from these figures that the turbulence was more severe near the exit than the entrance to the jets but this difference may not be significant because the number of cases was rather small and there were more flights made near exits than entrances to jets.

4. R.A.E. Turbulence Search Flights. During the period April to December 1951 the Aero. Flight of the Royal Aircraft Establishment, South Farnborough carried out some investigational flights on the same lines as the Royal Air Force "Upper Bumps" flights. Seven such flights to investigate turbulence near jet streams have been analysed by the same method and the results of the analysis are displayed in Figs. 8 and 9. A study of the most severe turbulence experienced in the four quadrants of the jet stream did not reveal anything of interest, mainly because the number of flights was small. None of the flights was made near the entrance or exit of the jet so it was not possible to learn anything from this series regarding the relative severity of turbulence in these two regions.

5. Summary and conclusions. 23 jet streams were investigated by the Royal Air Force and seven by the R.A.E. Aero. Flight. The worst turbulence encountered near any of the jets is shown in the following table:

Table II

| | R.A.F. Flights | R.A.E. Flights | Totals |
|---|---|---|---|
| Severe turbulence | 9 | 0 | 9 |
| Moderate turbulence | 9 | 1 | 10 |
| Perceptible turbulence | 3 | 6 | 9 |
| No turbulence | 2 | 0 | 2 |
| | 23 | 7 | 30 |

When examining the above table it must be realised that on account of various operational reasons, such as shortage of aircraft etc., the amount of flying carried out varied considerably with different jet streams. Moreover, in some of them the investigational flights may not have been made in the most turbulent part of the jet. There is no doubt that further work needs to be done before a true and comprehensive picture can be obtained but the following tentative conclusions can be put forward.

(a) All jet streams are not turbulent although the great majority of them are.

(b) When an aircraft is flying near a jet stream there is approximately an equal probability that the worst bumpiness encountered will be either severe, moderate or perceptible in intensity.

No decisive result was obtained from the study of the variation in the severity of the bumpiness with the horizontal and vertical shear in the vicinity of the jet stream (Figs. 3-7 and 9). But in general it appears that severe turbulence occurred when the vector difference between the wind at the 300 and 500 mb. level was 35 kt. or more and the horizontal shear was 0.2 hr.$^{-1}$ or more.

The analysis of all the investigational flights in relation to their position with respect to the axis of the jet, shown diagrammatically in Figs. 1, 2 and 8 confirms previous findings on the subject. With only a few exceptions the moderate and severe incidents occurred on the low pressure side of the jet and on the high pressure side above the axis; there were very few on the high pressure side below the axis.

An examination of the variation along the axis of the jet did not produce any really conclusive results.

## References

| | | |
|---|---|---|
| 1. | Hislop, G.S. | Clear Air Turbulence over Europe.<br>J.R. Aero. Soc. 55, 1951, p. 185. |
| 2. | Bannon, J.K. | Meteorological aspects of turbulence affecting aircraft at high altitude.<br>Prof. Notes, 7, No. 104, 1951. Meteorological Office, London. |
| 3. | Bannon, J.K. | Weather systems associated with some occasions of severe turbulence at high altitude.<br>Met. Mag., London, 81, 1952, p. 97. |
| 4. | Jones, D.C.E. | Weather systems associated with some occasions of severe turbulence in clear air at high altitude - a further analysis. M.R.P. 828. |

APPENDIX I

## REPORT OF OCCURRENCE OF CLEAR AIR TURBULENCE

1. Aircraft type and number — Canberra W.D. 986
2. Unit and Station — 617 Squadron Royal Air Force Binbrook
3. Captain and Crew — Fg.Off.Dowling Crew:- Fg.Offs.Johnson and White.
4. Date/time group in respect and incident 241235 - 241420z 24th April, 1952.
5. Position of start of turbulence See 11.
6. Instrument readings as at position given in 5 above:-
   - (i) Altimeter Sub-scale setting 1019 mb.
   - (ii) Indicated Airspeed )
   - (iii) Magnetic Heading )
   - (iv) Indicated Air Temperature )
7. Magnitude of turbulence ) See 11. below
8. Extent and depth of turbulent region )
9. Weather conditions in locality of incident )
10. Other information )
11. Remarks. Upper Bumps. Special flight to investigate jet stream extending from the Western Isles to Wales to Brittany. Aircraft sent on route Binbrook - Castle Archdale - Binbrook.

| Time (GMT) | Position (Lat° | Long°) | Turbulence Commenced (Height in ft.) | Finished | I.A.S. (Kt.) | Heading (°Mag.) | Indicated Temperature (°C) |
|---|---|---|---|---|---|---|---|
| 1242 | 53.25N | 01.10W | Slight. 21,000 | 25,000 | 300 | 299 | - |
| 1247 | 53.35N | 02.10W | Slight. 35,000 | for 6 mins. | 250 | 299 | -27 |
| 1255 | 53.47N | 03.20W | Slight. 32,500<br>(None above 35,000) | 35,000 | 230 | 299 | -28/-27 |
| 1306 | 53.55N | 05.12W | Slight. 35,000<br>(Increased slightly at 33,000. Very slight to 24,000) | 31,000 | 230 | 310 | -34 |
| 1313 | 54.09N | 06.00W | Slight. 25,000 | 34,500 | 230 | 310 | -31 at 34,000 |
| 1315 | 54.12N | 06.15W | Moderate 29,000 (moderate at this level only.) | | | | |
| 1326 | 54.30N | 07.25W | Slight. 34,500<br>(Moderate at 39,000 only) | 41,500 | 230 | 108 | -30 |
| 1336 | 54.21N | 05.27W | Mod. bec. Slight 39,000 | 41,500 | 230 | 115 | -31 |
| 1343 | 54.03N | 04.15W | Slight 38,000<br>(Moderate at 24,000 only) | 23,000 | 230 | 110 | - |
| 1352 | 53.30N | 02.12W | Moderate 24,000<br>(Extended at this level to 53.29N 01.59W.) | - | 230 | 110 | -24 |
| 1355 | 53.25N | 01.24W | Slight 29,500 | 31,000 | 230 | 110 | -23 |
| 1359)<br>1402) | Descent at Binbrook | | Moderate 16,000 | - | 230 | 075 | - |

Weather over route Generally cloudy with cloud tops at 9,500 feet.
Beyond 5deg. W. Rolls of Cirrus at 31,000/34,000 feet.
Around 7deg. W. Rolls of Ac/As at 27,500 feet.

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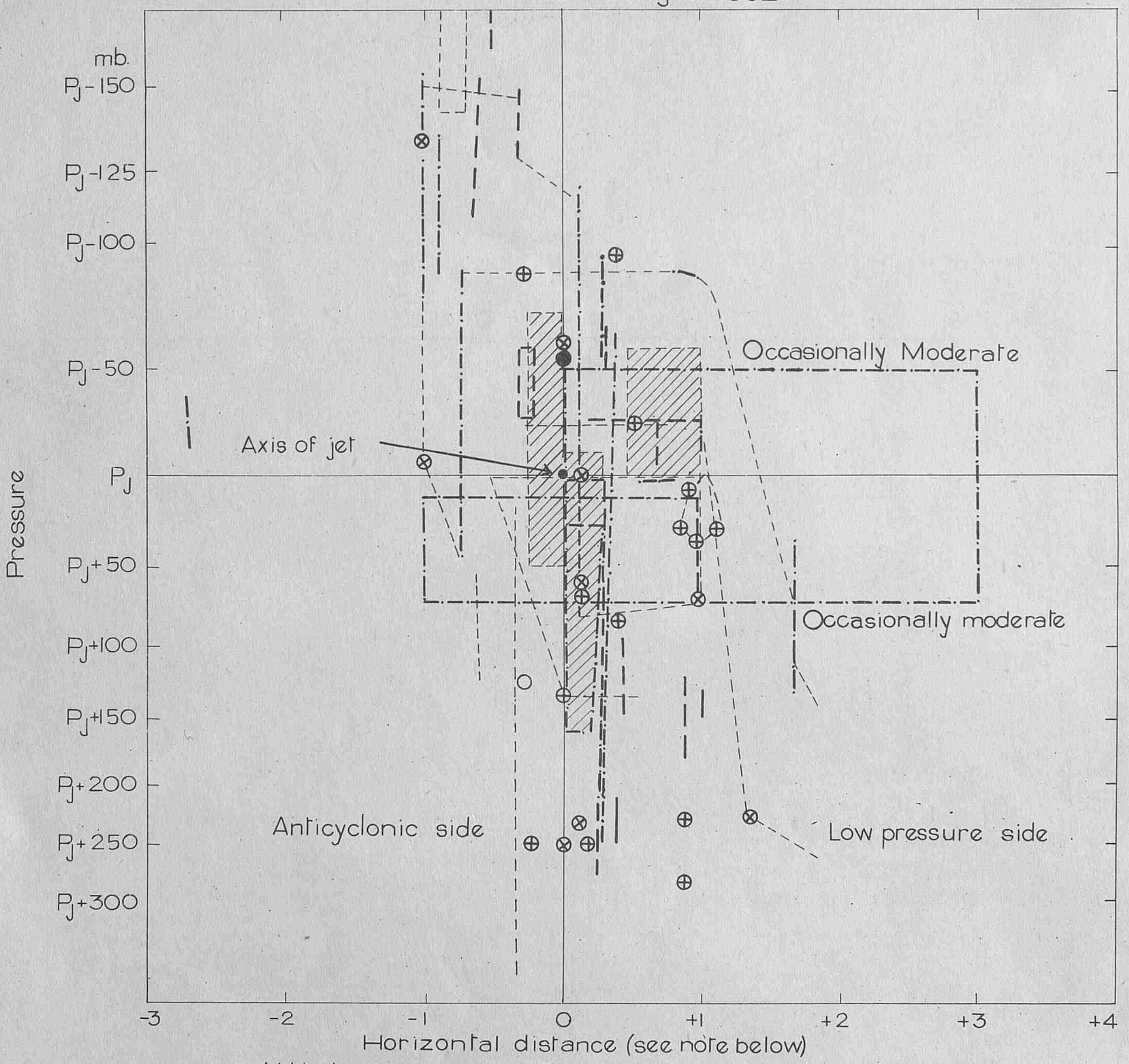
Fig. 1

Positions relative to the axis of jet streams of flights made by the R.A.F.

14 jet streams

March-August 1952



○ or _ _ _ Nil turbulence ⊕ or _ . _ . _ Perceptible turbulence ⊗ or _ _ Moderate turbulence

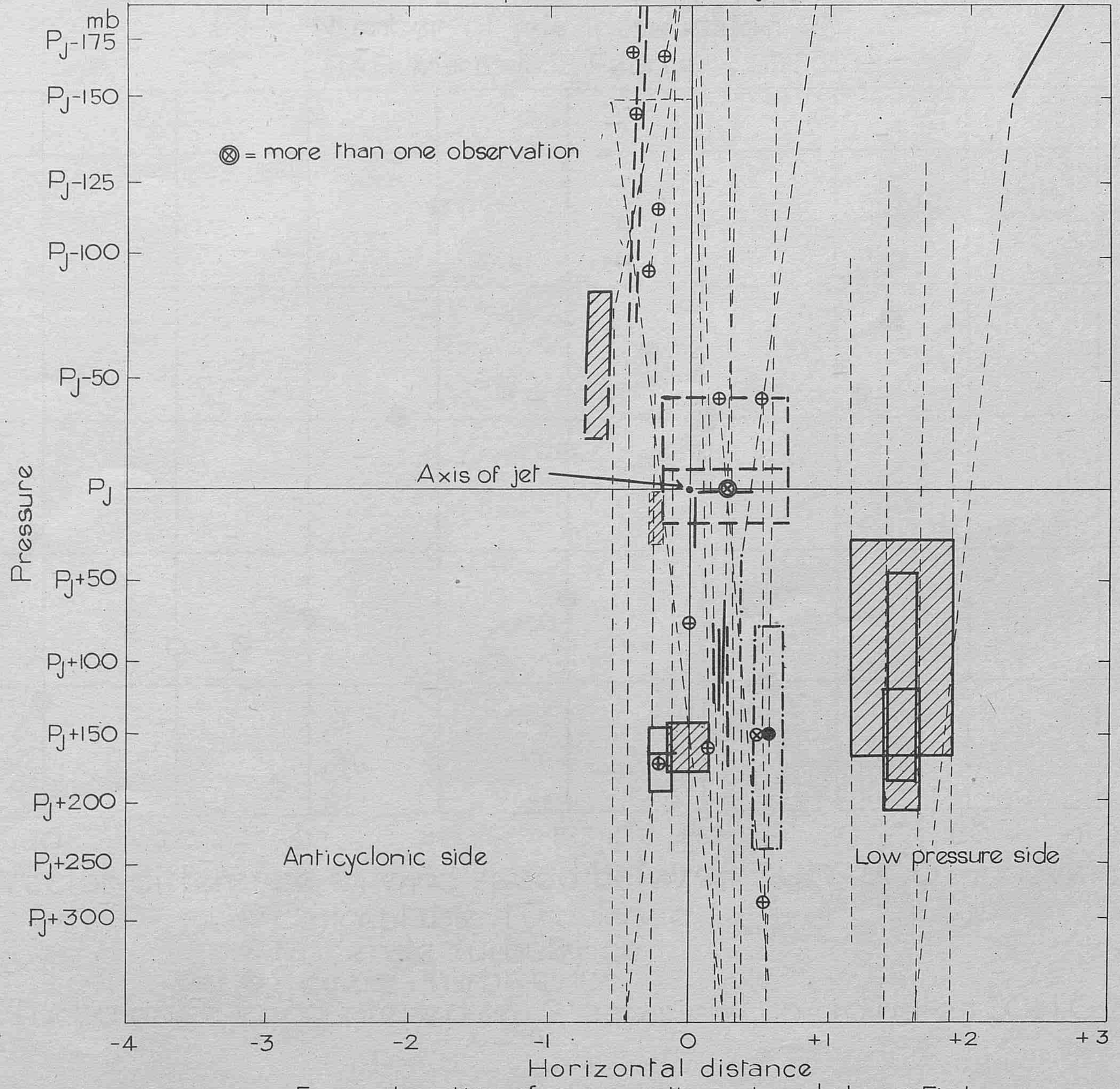
● or ___ Severe turbulence ▨ Zone of turbulence

$P_J$ is the pressure at the axis of the jet

Note:- The unit of horizontal distance is the distance on the low pressure side of the jet axis in which the wind speed is half its maximum value

Fig.2

Positions relative to the axis of jet streams of special flights made by the R.A.F.
9 jet streams
September 1952 - February 1953



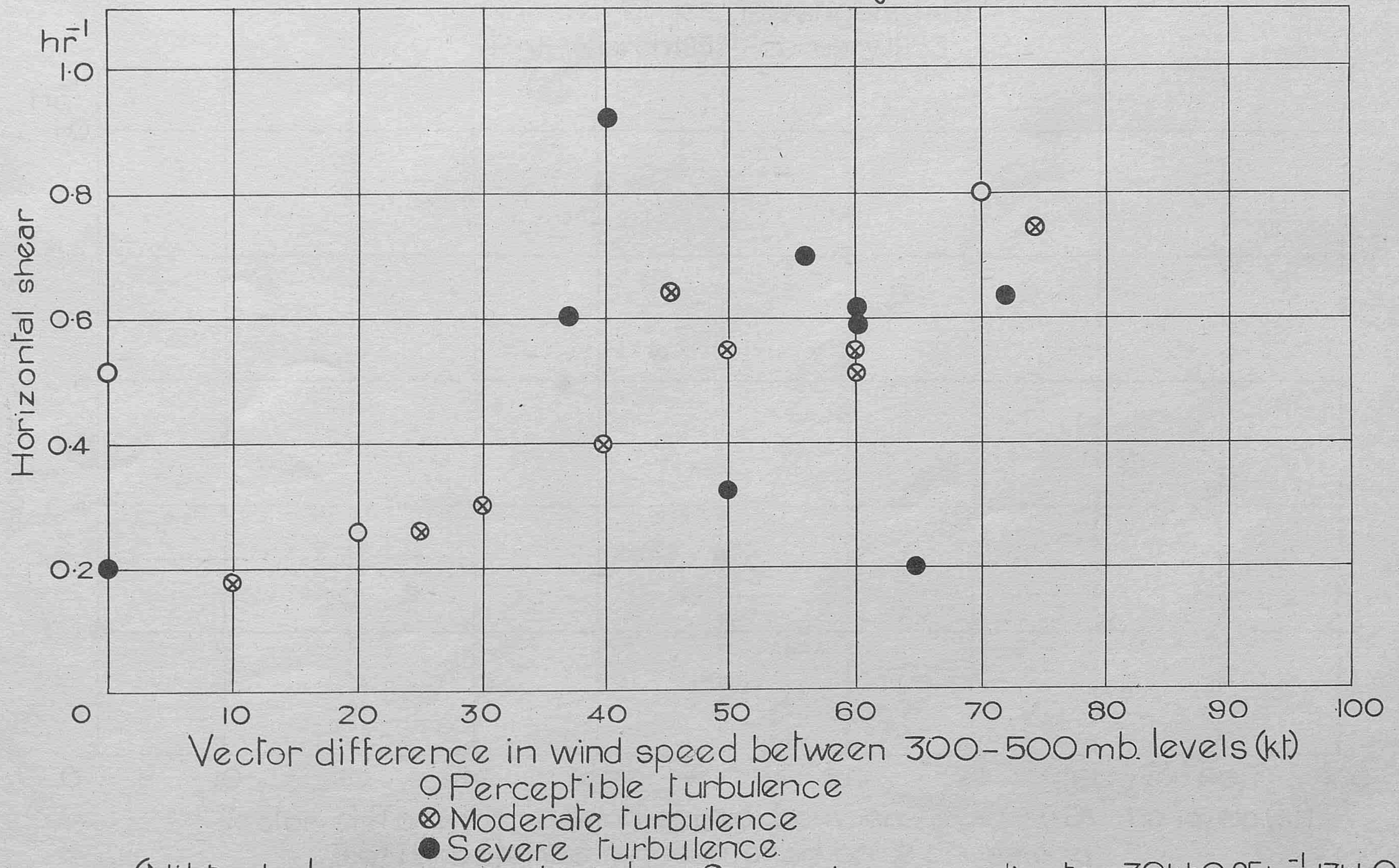
For explanation of axes, units and symbols see Fig.1

Fig.3

Most severe turbulence encountered in any quadrant of a jet stream

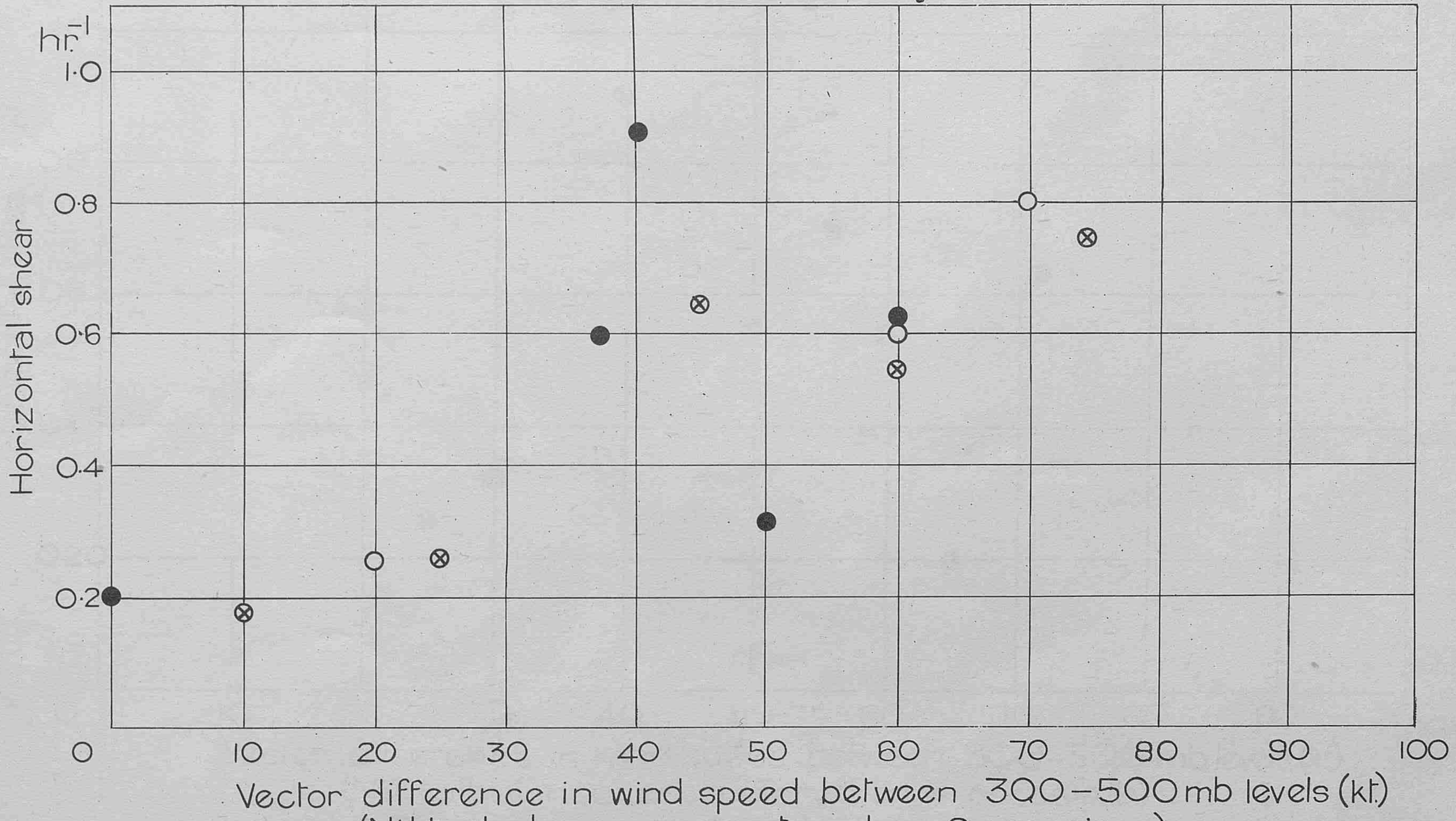
Number of jets investigated-23

R.A.F. March 1952-February 1953



(Nil turbulence encountered on 2 occasions, co-ordinates 30kt. 0·25hr.$^{-1}$; 17kt 0·2hr.$^{-1}$)

Fig. 4
Most severe turbulence encountered in the low pressure side below axis of jet
Number of jets investigated = 15
RAF. March 1952 - February 1953



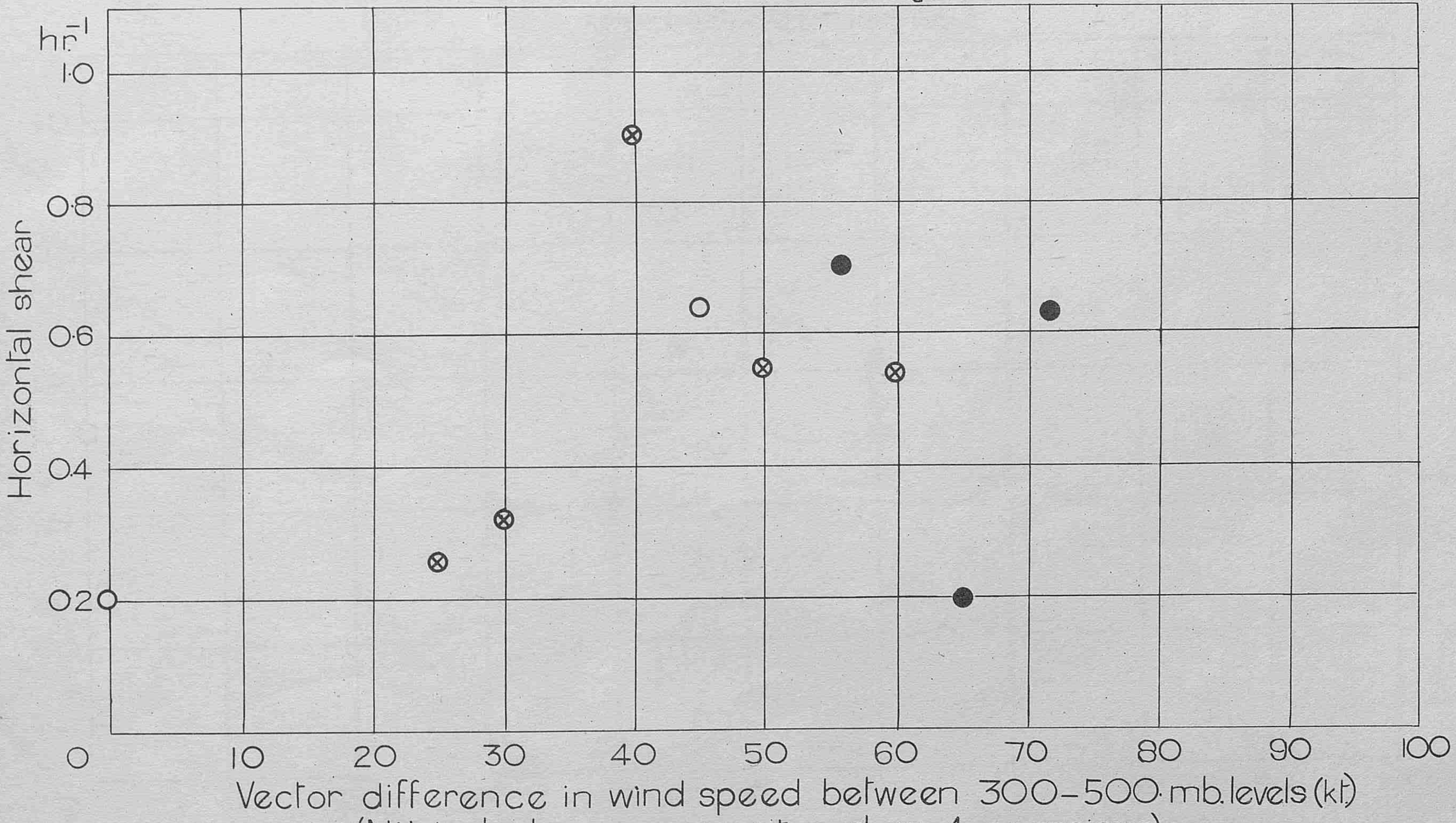
(Nil turbulence encountered on 2 occasions)
For explanation of symbols see Fig. 3

Fig. 5

Most severe turbulence encountered in the low pressure side above axis of jet

Number of jets investigated = 13

R.A.F. March 1952-February 1953



(Nil turbulence encountered on 4 occasions)

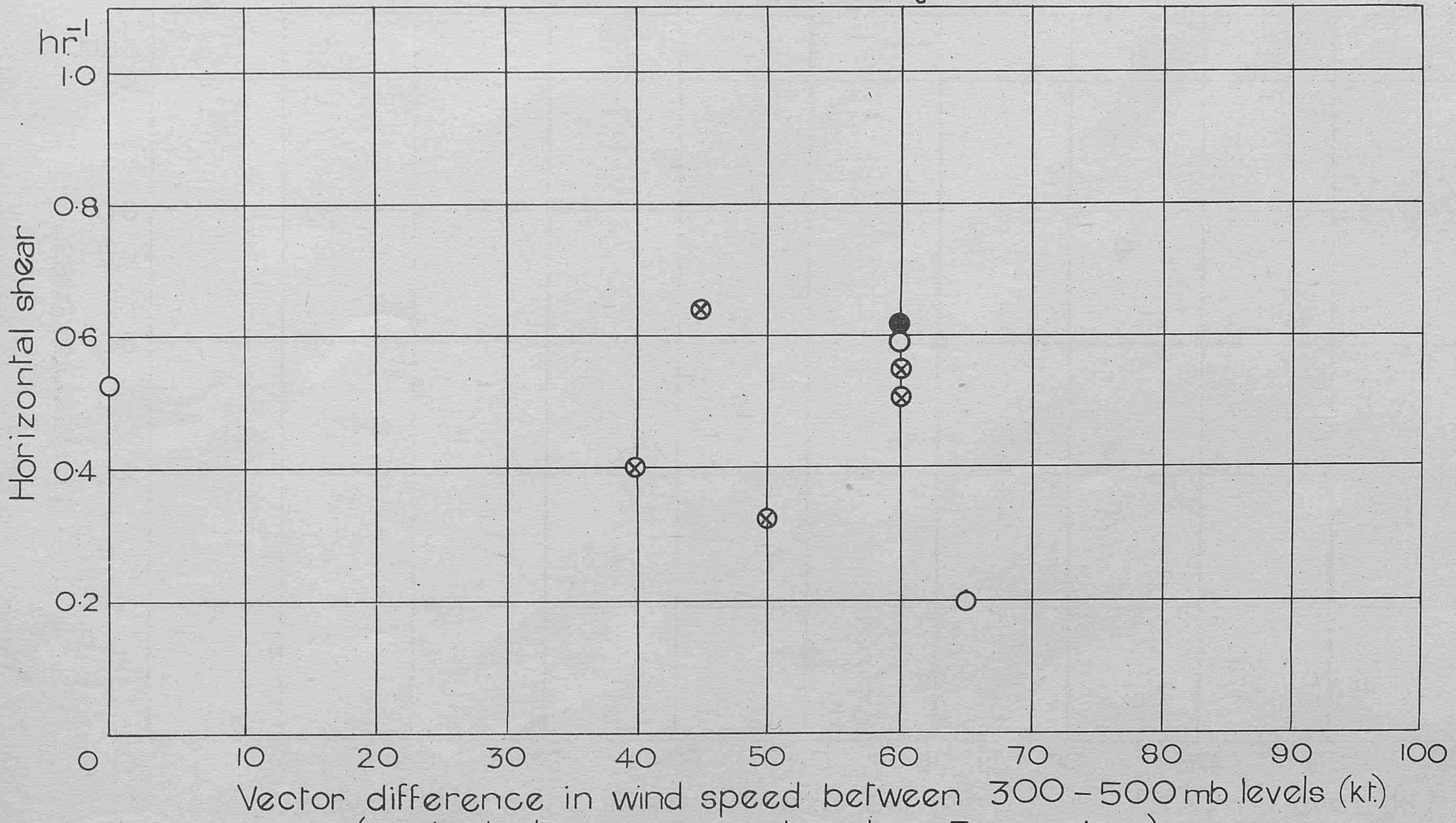
For explanation of symbols see Fig. 3

Fig. 6

Most severe turbulence encountered in the high pressure side above axis of jet

Number of jets investigated=12

R.A.F. March 1952-February 1953



(Nil turbulence encountered on 3 occasions)

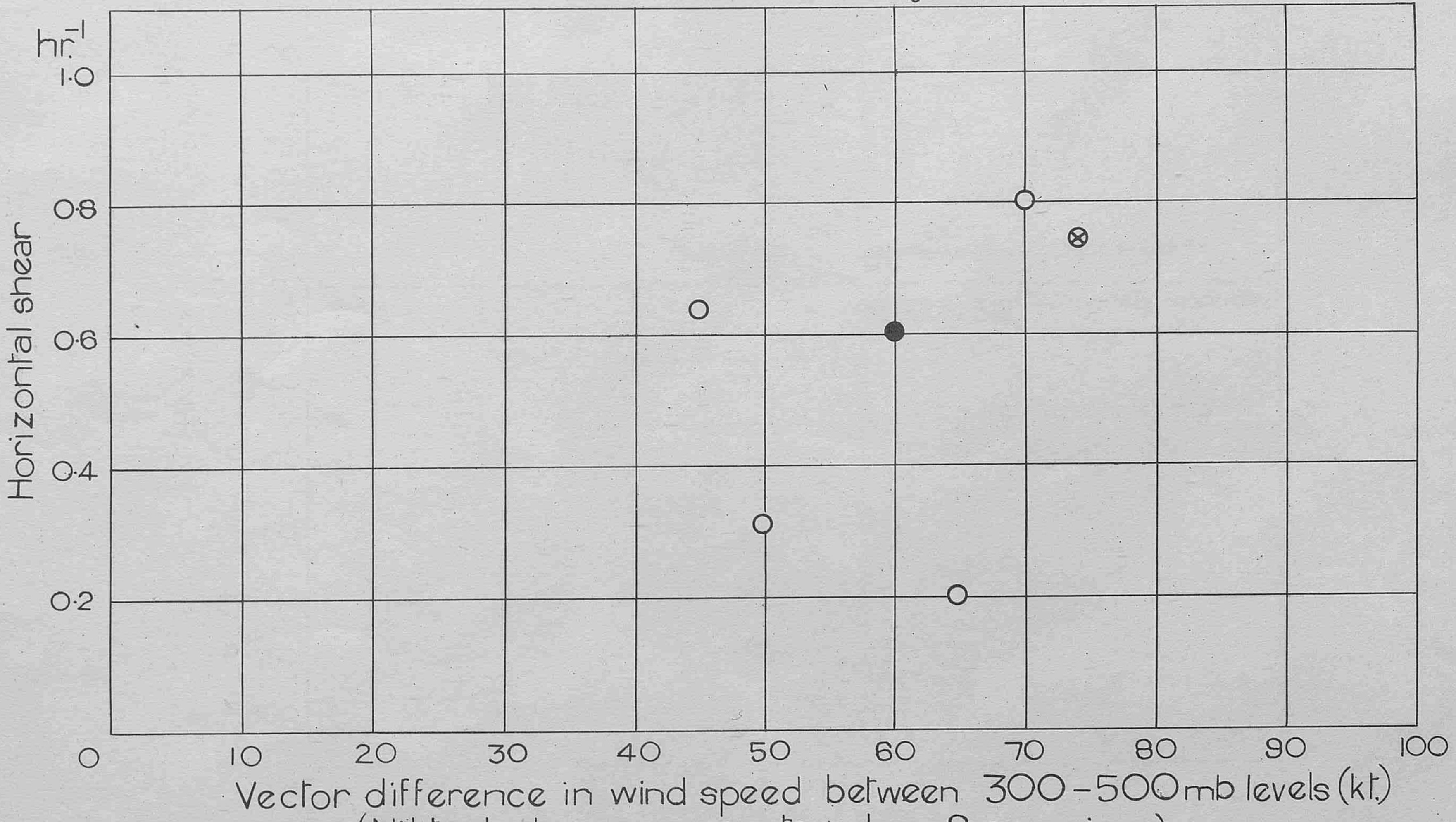
For explanation of symbols see Fig. 3

Fig. 7

Most severe turbulence encountered in the high pressure side below axis of jet

Number of jets investigated=14

R.A.F. March 1952-February 1953



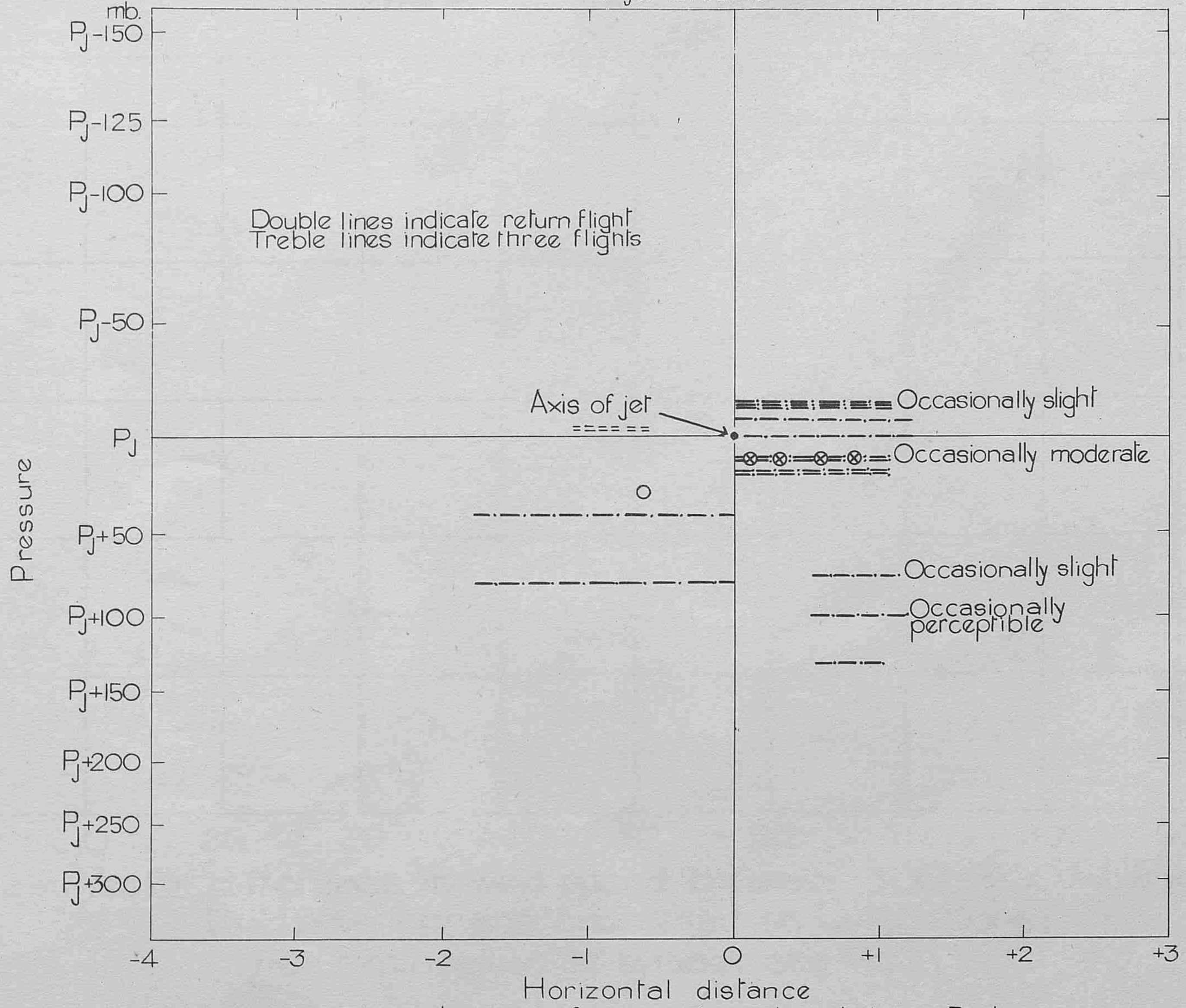
(Nil turbulence encountered on 8 occasions)

For explanation of symbols see Fig. 3

Fig. 8

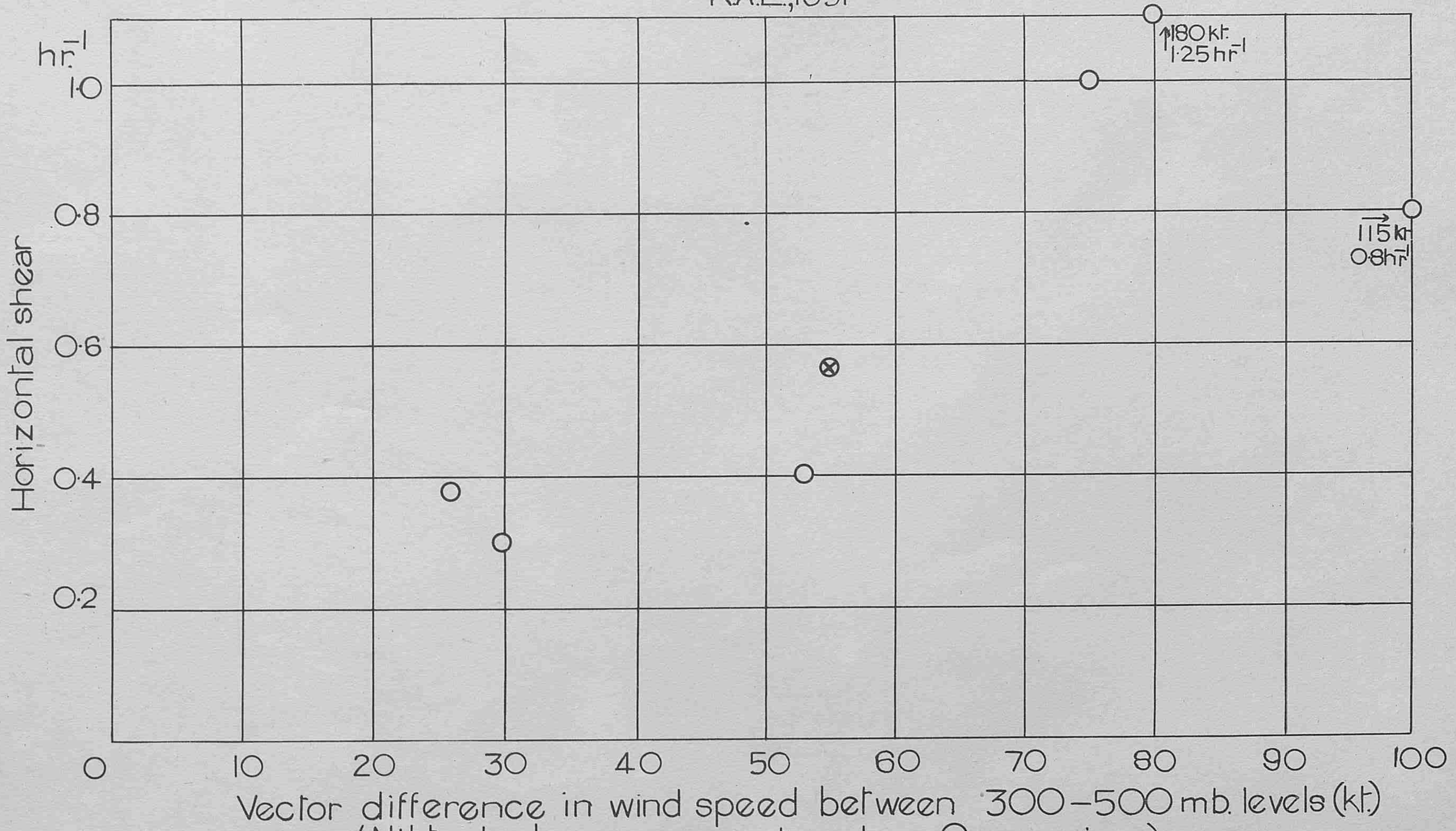
Positions relative to the axis of jet streams of special flights made by Aero. Flight R.A.E.

7 jet streams



For explanation of axes units and symbols see Fig. 1

Fig. 9
Most severe turbulence encountered in any quadrant near a jet stream
Number of jets investigated = 7
R.A.E., 1951



(Nil turbulence encountered on 0 occasions)
For explanation of symbols see Fig. 3.

MRP's 828, 829, 830, 831, 833

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 828 by Mr. D.C.E. Jones
A copy is available in the Library of the Meteorological Office."

M.R.P. 828

S.C.III/160

24th July, 1953.

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METEOROLOGICAL RESEARCH COMMITTEE

# Weather systems associated with some occasions of severe turbulence in clear air at high altitude: a further analysis

by

D.C.E. Jones (Meteorological Office)

## Summary

This paper is an extension of work carried out by Bannon (1,2) and largely confirms the results which he found.

A further 147 reports of severe turbulence encountered in clear air at altitudes between 15,000 and 48,000 ft. over or near the British Isles were analysed in relation to broad features of the 300 mb. contour chart. Seventy-one per cent of the occasions were associated with a jet stream, nine per cent were not associated with any particular feature of the charts and nearly all the remaining twenty per cent were associated with upper lows, upper troughs or strong winds. The occurrences near jet streams were again found to be grouped mainly on the low pressure side of the jet and to a lesser extent above the axis on the anticyclonic side. Less than ten per cent were found to be below the jet axis on the high pressure side.

The highest level at which severe turbulence was encountered was 48,000 feet.

1. Introduction. In previous papers (1,2) attempts have been made to correlate the occurrence of severe turbulence with various meteorological situations and it was found that the large majority was in the vicinity of jet streams.

This paper describes a further analysis of all reports of turbulence encountered between 15,000 and 48,000 ft. and which were described as "severe" or "very severe" during the period August, 1951, and February, 1953. It was considered that reports of a lesser severity, (that is "moderate", "slight", "perceptible", etc.) were not of any great importance in an investigation of this sort and therefore only the "severe" cases were taken into consideration.

2. Method of reporting. All the reports of turbulence received were qualitative assessments because none of the aircraft participating in this investigation carried an accelerometer. Most of the reports were received from operational Royal Air Force jet bombers and fighters but some were also received from Royal Navy and Civil aircraft.

Since all the reports are qualitative it is difficult to be certain whether all pilots adopted the same standard of severity when reporting. There is always the risk that a pilot may describe an occurrence as severe because he considered it severe compared with the usual turbulence at high levels and not on an absolute scale. In order to try to overcome this possible human error and in an endeavour to achieve consistency and uniformity, the following statement was included on all forms used for entering reports on incidents of turbulence:

> "The following may serve as a rough guide in the assessment of the turbulence encountered:-

/(a)

(a) Perceptible (± .05g to ± .2g) - Perceptible difference from steady flying conditions.

(b) Moderate (± .2g to ± .5g) - Uncomfortable, slight tendency to be lifted out of seat.

(c) Severe (± .5g upwards) - Difficulty in observing flying instruments and/or maintenance of aircraft heading. Marked tendency to be thrown from seat."

Some of the incidents reported were unusually severe and two examples are given below:-

(a) "Aircraft (Meteor) was pitching and tossing about the sky and one of its instruments was shaken from the instrument panel."

(b) "Crew were lifted from their seats. Aircraft (Canberra) suffered structural damage and several rivets were dislodged."

Other reports described the turbulence as "moderate to severe" and "mainly moderate but occasionally severe" etc. but for the purpose of this analysis were regarded as "severe".

In all, 147 reports obtained during the period August, 1951, to February, 1953 were studied. The true airspeed of the aircraft at the time when the turbulence was encountered varied between about 250 and 450 knots.

3. Method of analysis. Each report was studied in conjuction with the information given in the Daily Aerological Record and an attempt was made to associate the turbulence with broad features of the relevant 300mb. contour chart, that is, with jet streams, strong winds, lows, troughs and discontinuity in the tropopause surface. A summary of the cases occurring in various weather situations is given in Table I below.

### Table I

### Weather situations associated with occasions of severe turbulence in clear air at high altitude

| Weather Situation | No. of Cases |
|---|---|
| Near jet stream | 98 |
| Probably near jet stream (i.e. jet stream present but details of maximum wind speed, height of jet etc. not available) | 7 |
| Upper trough | 19 |
| Upper low | 3 |
| Discontinuity in tropopause surface at same level as turbulence and within 100 miles of it. | 1 |
| As above and near upper trough | 2 |
| As above and near upper low | 0 |
| Strong upper winds but no evidence of a well defined jet stream | 4 |
| Unclassified | 13 |
| Total | 147 |

The thirteen unclassified cases occurred in situations that could not be classed as any one of the above types. A few of them were in the middle of upper ridges and there was no vertical or horizontal wind shear, nor a discontinuity of any kind that could be detected.

/Of

### Table II

### Mean values and extremes of various parameters of jet streams in which 98 severe turbulence incidents were reported

| | Wind Speed at the axis (knots) | Pressure at the axis (mb.) | Horizontal scale (n. mi.) |
|---|---|---|---|
| August, 1951 to August, 1952 | | | |
| Mean | 115 | 283 | 147 |
| Extremes | 80 (minimum)<br>170 (maximum) | - | 50 (minimum)<br>300 (maximum) |
| September, 1952 to February, 1953 | | | |
| Mean | 122 | 315 | 162 |
| Extremes | 87 (minimum)<br>165 (maximum) | - | 70 (minimum)<br>360 (maximum) |

The horizontal unit is the distance from the axis, at the same level on the low pressure side, at which the wind speed is half the value at the axis.

As in the previous analysis the most striking result is that severe turbulence was encountered on less than ten per cent of occasions in the quadrant below and on the high pressure side of the axis, and 75 per cent were in the two quadrants on the low pressure side.

An attempt was made to study any variations with respect to position along the jet stream. But only three cases occurred near the entrance and eleven near the exit of a jet and these numbers are too small for any particular significance to be attached to them.

5. Conclusions. This analysis of occasions of severe turbulence occurring over and near the British Isles in clear air at high levels confirms a previous finding that the major number occur in association with well defined jet streams and that most of the others were associated with upper lows, upper troughs or strong winds. About nine per cent appeared to have no obvious association with any pattern in the upper air. 75 per cent of the cases reported near jet streams were on the low pressure side of the jet. Of the remaining 25 per cent (on the high pressure side) less than ten per cent were below the axis.

Little or no further progress has been made in this investigation towards formulating any rigid rules to assist the practical forecaster in his task of predicting the occurrence of high level turbulence in clear air. Turbulence is a state in which the air undergoes rapid and random local fluctuations and it appears very probable, at this stage, that it is only the statistical properties that can be recognised and subjected to analysis. Recent reports from jet aircraft indicate that severe turbulence can be experienced in practically any weather situation up to a height of at least 48,000 feet and well above the tropopause. But, nevertheless, the results obtained in these investigations should be of some help to the forecaster in recognising the regions of the atmosphere where aircraft flying at speeds of between 250 and 450 knots stand the greater risk of encountering patches of severe bumpiness.

### References

1. Bannon, J.K. Severe turbulence encountered by aircraft near jet streams. Met. Mag., London, 80, 1951, p.262.

2. Bannon, J.K. Weather systems associated with some occasions of severe turbulence at high altitude. Met. Mag., London, 81, 1952, p.97.

G.376175/JWP/7/53/65.

Of the 147 cases, 21 occurred in the stratosphere, two were almost at the tropopause level and the remainder were in the upper troposphere.

As regards the height at which the severe turbulence was encountered, the majority of cases were between 25,000 and 35,000 ft., 25 between 35,000 and 40,000 ft., 7 between 40,000 and 45,000 ft. and one was above 45,000ft. The latter was experienced in a Canberra over Leicester on 19th October, 1951, between 47,500 and 48,000 ft. It occurred near an upper trough and about 10,000 feet above the tropopause.

It is not safe to conclude from the above figures that the incidence is less in the lower stratosphere because in all probability the amount of flying at those levels was considerably less than at those below 40,000 ft.

4. Jet Streams. The reports of severe turbulence in the vicinity of jet streams were further analysed with regard to their position relative to the axis of the jet. The method used was the same as that employed in previous work (1,2). By inspection of the 300 mb. charts and the wind reports given in the Daily Aerological Record the position, orientation and height of the axis of the jet were estimated at the time the turbulence was reported. For this purpose it was assumed that the jet stream was defined by more or less straight and parallel contours with a maximum wind speed of 80 knots or more, and the contour gradient decreasing on both sides of the maximum.

The level of the jet stream was taken as the level of the maximum wind reported on the upper air soundings in the vicinity of the jet. This wind report was taken as the nearest approximation to the maximum speed of the jet.

When a report of turbulence did not coincide with the time of routine upper air soundings an allowance was made for movement of the jet axis by using continuity considerations of space and time.

As in previous work (1,2) the horizontal unit of measurement was taken as the distance from the centre of the jet stream, on the low pressure side, over which the wind velocity at the same level as the axis of the jet fell to half the maximum value. The vertical co-ordinate was taken as the difference between the I.C.A.N. pressure equivalent of the height and the pressure at the axis of the jet.

This method of analysis has the following disadvantages:-

(a) The jet axis cannot be positioned very accurately but on the other hand the direction of the jet must be reasonably accurate on most occasions.

(b) The height of the jet axis need not necessarily be the same as the height of the wind maximum observed at a point near the jet but, in general, the difference should be small. Bearing in mind that in the normal jet stream the isopleths of wind speed are relatively close together in the quadrants below the axis on the low pressure side and above the axis on the high pressure side, it follows that the height of the axis was probably underestimated for soundings on the low pressure side and overestimated for those on the high pressure side.

(c) An analysis of this kind must be subjective to a certain extent.

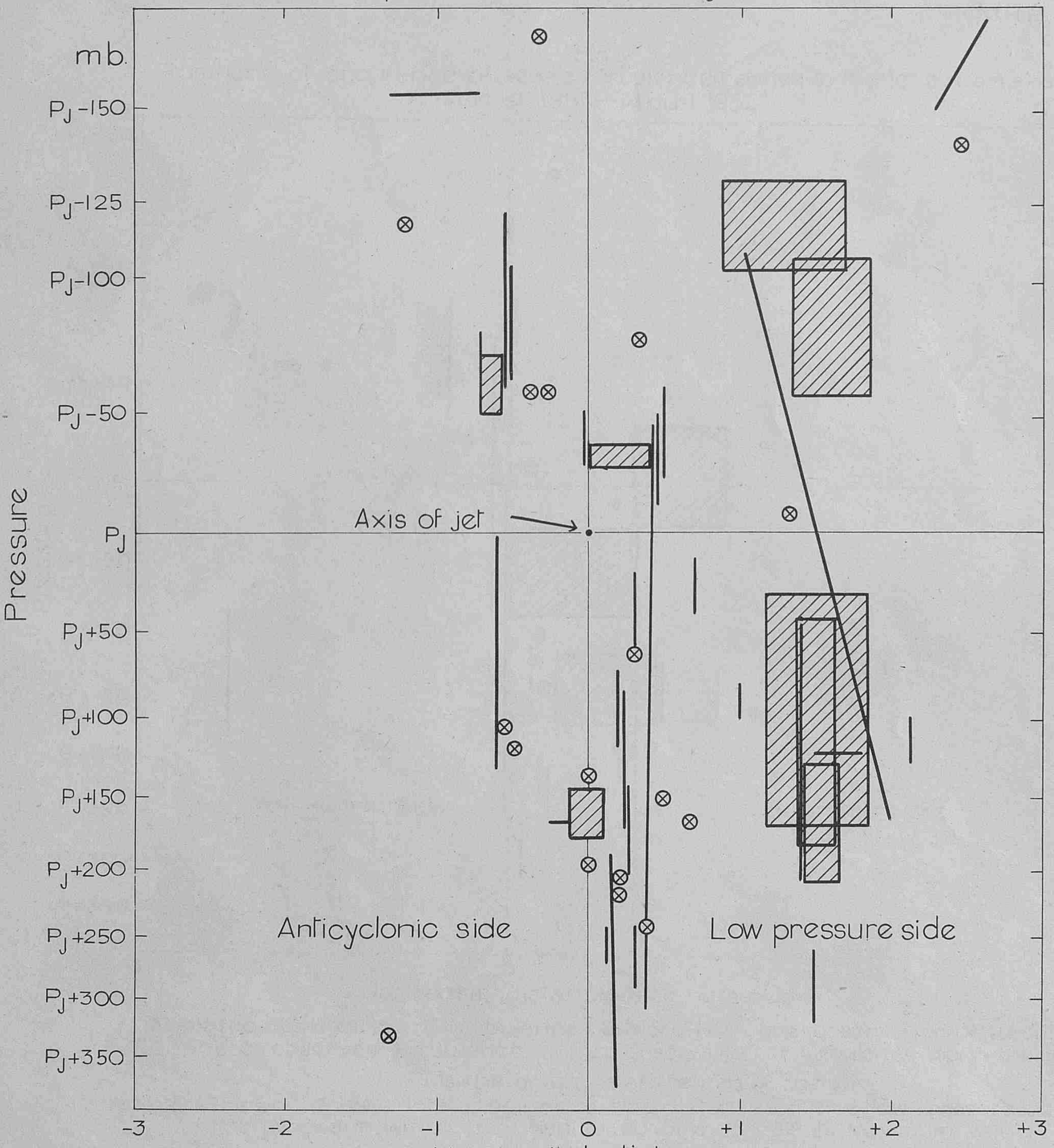
It is claimed, however, in view of the fairly large number of cases examined that the accuracy obtained is sufficient to show the main features, although there may be some uncertainty in any particular case. As already indicated in Table I 98 cases were deemed to have been near jet streams. A further 7 cases were probably within the sphere of influence of a jet stream but there were not enough data available to make a satisfactory assessment of the position, height and maximum wind at the axis of the jet.

The co-ordinates thus obtained for the positions of the reports with respect to the jet stream were plotted in Figs. 1 and 2 which show all the cases examined for the periods August, 1951, to August, 1952 and September, 1952 to February, 1953 respectively. Details regarding certain parameters of the jet streams are given in Table II below.

/Table II

Fig. 2

Positions of occurence of severe turbulence relative to the jet stream axis
September 1952 – February 1953



For explanation of axes and symbols see Fig. 1

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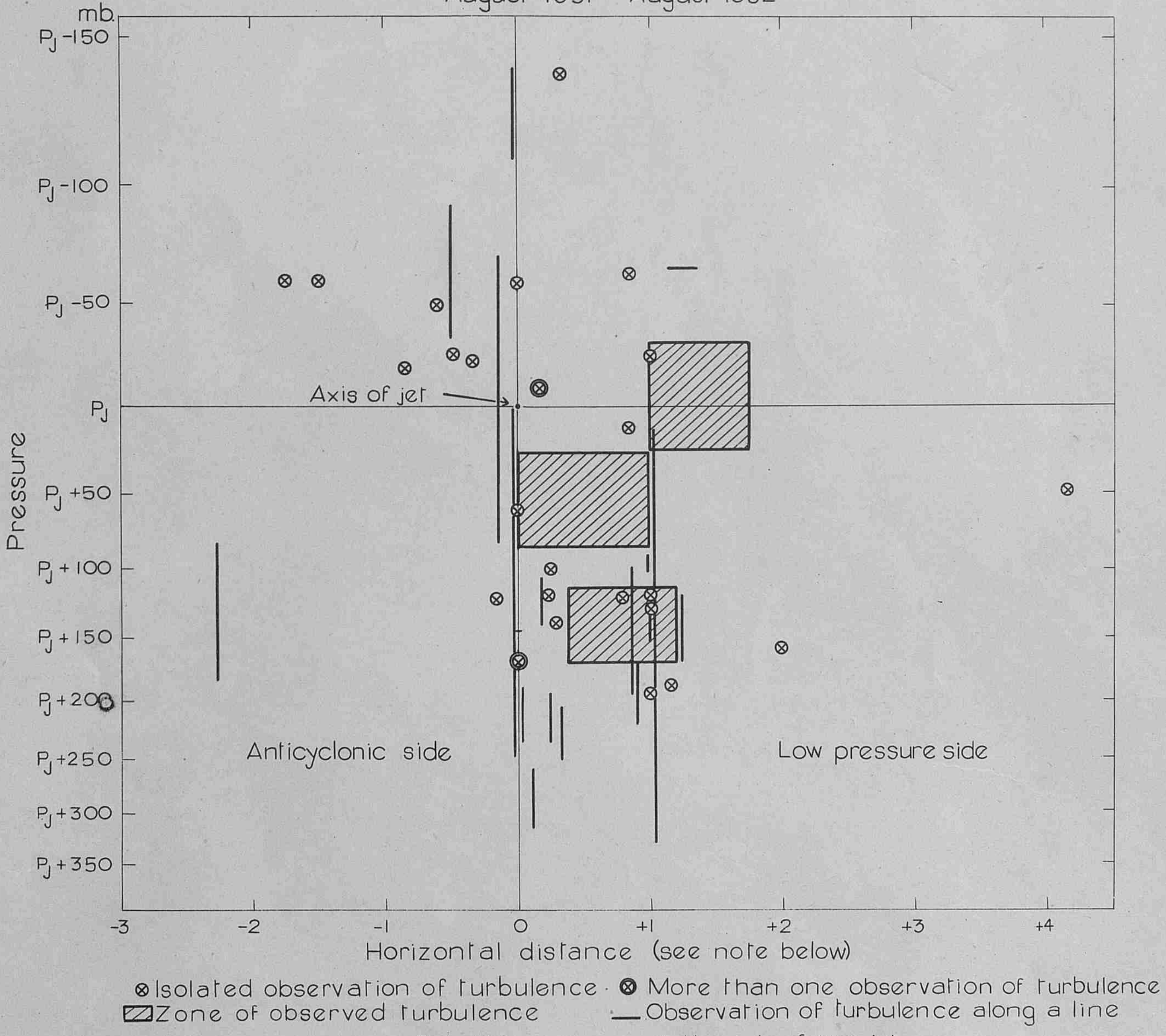
Fig. 1
Positions of occurence of severe turbulence relative to the jet stream axis
August 1951 — August 1952



$P_J$ is the pressure at the axis of the jet

Note:- The unit of horizontal distance is the distance on the low pressure side of the jet axis in which the wind speed falls to half its maximum value

C.D. 8053

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M.R.P. 829
S.C.II./157
6th August, 1953.

AD 39445

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

# Vorticity and Synoptic Development

by

J.S. Farquharson (Meteorological Office)

## 1. Purpose.

In the work of the Forecast Division of the Meteorological Office particular attention is given to published research or other available knowledge which offers any prospect of an improvement in the standard of forecasting. Over some years now such attention has been given to the practical application of the ideas of Sutcliffe (1947) and of new techniques, based on them, suggested by Sutcliffe and Forsdyke (1950). Experience has shown that while illuminating and valuable as a background in synoptic analysis and prognosis, nevertheless these ideas in practice sometimes confront the forecaster with puzzling discrepancies, as when, for example, pressure rises over a wide area covered by a cold trough in the upper air. It is the purpose of this paper first to illustrate difficulties, then to suggest possible reasons for them and finally to extend the discussion of the scalar vorticity equation.

## 2. The Sutcliffe Development Term

Sutcliffe (loc. cit.) derived an expression for the relative divergence between pressure levels as follows:

$$f\,\mathrm{div}\,V - f\,\mathrm{div}\,V_0 = -V'\frac{\partial}{\partial s}(\zeta + \zeta_0 + f) \qquad (1)$$

where $f$ is the Coriolis term, $V$ and $V_0$ are wind vectors at two pressure levels in the atmosphere, $V'$ is the thermal wind speed between these two levels, $\partial/\partial s$ represents differentiation along the thermal wind and $\zeta$ and $\zeta_0$ are the vorticities at the two levels.

This expression is taken as a measure of synoptic development, following the method of the atmosphere, and Sutcliffe suggests that the levels of 1000 mb. and 500 mb. may be used, though disclaiming any particular merit in that choice.

As a matter of interest it was thought desirable to consider the effect of varying the thickness. Accordingly, for 4th February, 1951, when a depression of some fair intensity affected southern Ireland the development given by equation (1) was computed for each of the thicknesses 1000-700 mb., 1000-500 mb., and 1000-300 mb. and for each of the hours 03,09, 15 and 21 G.M.T. - the method

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Sutcliffe (loc. cit.) derived an expression for the relative divergence between pressure levels as follows:

$$l\,\mathrm{div}_p \underline{V} - l\,\mathrm{div}_p \underline{V}_0 = -V'\frac{\partial}{\partial s}(l+\xi+\xi_0) \qquad (1)$$

where $l$ is the Coriolis term, $\underline{V}$ and $\underline{V}_0$ are wind vectors at two pressure levels in the atmosphere, $V'$ is the thermal wind speed between these two levels, $\partial/\partial s$ represents differentiation along the thermal wind and $\xi$ and $\zeta_0$ are the vorticities at the two levels.

This expression is taken as a measure of synoptic development, following the motion of the atmosphere, and Sutcliffe suggests that the levels of 1000 mb. and 500 mb. may be used, though disclaiming any particular merit in that choice.

As a matter of interest it was thought desirable to consider the effect of varying the thicknesses. Accordingly, for 4th February, 1951, when a depression of very rare intensity affected southern Ireland the development given by equation (1) was computed for each of the thicknesses 1000-700 mb., 1000-500 mb., and 1000-300 mb. and for each of the hours 03, 09, 15 and 21 G.M.T. - the method of Matthewman and Sawyer (1951) being employed. Isopleths of development were constructed, the unit being $10^{-2}$ hour$^{-2}$, and these are shown in Figures 1 to 12. The isopleths are shown by solid lines, the relevant surface isobars by broken lines, and, on the 1000-500 mb. charts, the thickness isopleths of that thickness by dotted lines. Cyclonic development is shown by negative values of the development.

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/Referring

Ind #2

11a

Referring first to the charts for 0300 G.M.T., we see that the depression west of Ireland was located at the northern tip of a warm tongue in the 1000-500 mb. thickness pattern, with a developing cold trough to the southwest of it. The area of maximum cyclonic development was apparently well in the cold air, with a subsidiary such maximum over the front, over southwest Ireland. On the 1000-700 mb. thickness chart there were similar centres of maximum cyclonic development in roughly similar areas; they were of about the same intensity except that the one over southwest Ireland was apparently of greater extent. On the 1000-300 mb. thickness chart the cyclonic maxima are increased relative to those on the 1000-500 mb. chart, and, while that over southwest Ireland is again broadly in the same location as those on the other two charts, the one on the Atlantic is appreciably nearer the centre of the depression than the corresponding one at 500 mb.

By 0900 G.M.T. the centres of the depression had moved in over western Ireland, the lowest pressure having fallen by two millibars. On the 1000-700 thickness chart the area of maximum cyclonic development is near the Isle of Man and again in proximity to the major front of the depression. The subsidiary maximum on the Atlantic has become much less intense but more extensive. On the 1000-500 mb. chart the area of maximum cyclonic development is over northwest Ireland, almost over the main centre of the depression, and its value is similar to that shown on the Atlantic on the earlier corresponding chart. On the 1000-300 mb. chart the area of maximum cyclonic development is appreciably further west than the corresponding area on the 1000-500 mb. chart, its value being similar to that for the same thickness six hours earlier.

At 1500 G.M.T. the depression was centred over the heart of Ireland, its central pressure being lower than 944 mb. Apparently the tendency for a double centre was dying out. The thickness lines on the 1000-500 mb. chart show a pool of cold air south of the centre of the low. The 1000-700 mb. development chart shows what is apparently a decaying cyclonic maximum over northeast Ireland with an anticyclonic maximum off southwest Ireland. The corresponding chart for 1000-500 mb. shows an area of cyclonic development from north of Stornoway southwards over Ireland and another to the southwest of Ireland. The intensity of this development is much less than the maximum of six hours earlier. On the 1000-300 mb. development chart the area of maximum cyclonic development is off northwest Ireland and, again, its intensity is reduced relative to the corresponding maximum of six hours earlier. As on the thickness chart for 1000-500 mb., there is a subsidiary maximum of cyclonic development off southwest Ireland, though further west and north.

By 2100 G.M.T. the depression was centred over eastern Ireland and the Irish Sea and the area of maximum cyclonic development on the 1000-700 mb. chart was just ahead of it over northern England. On both the charts for 1000-500 mb. and 1000-300 mb., however, the area of maximum cyclonic development is shown to the northwest of the depression, apparently in the relatively cold air. This tendency for the area of maximum cyclonic development to retrogress into the cold air as the thickness dealt with is increased is shown on all the charts except those for 0300 G.M.T. and even on these the conclusion must be somewhat doubtful. We see then that the region of maximum cyclonic development varies with the thickness of the atmosphere considered.

## Restrictive Conditions.

McVittie (1952) has dealt at length with the restrictive conditions which apply in the derivation of the development term but it may be of interest to proceed from Sutcliffe's own equations to see where we arrive.

To obtain equation (1), Sutcliffe (loc. cit.), on the assumption that the sum of

$$-\xi \,\mathrm{div}_p \underline{V} - \frac{\partial v}{\partial p}\left(\frac{\partial}{\partial x}\frac{dp}{dt}\right)_p + \frac{\partial u}{\partial p}\left(\frac{\partial}{\partial y}\frac{dp}{dt}\right)_p$$

is negligible derives the approximate scalar vorticity equation:

$$\frac{d}{dt}\left(\xi + l\right) = -l \,\mathrm{div}_p \underline{V} \qquad \ldots\ldots\ldots\ldots (2)$$

Eady (1952) says: "This is certainly not true in the vicinity of a "jet" and the relative error may also be large near intense depressions or anticyclones. In general the errors are larger at high than at low levels ........". It may be that these words of Eady's provide the explanation for the apparent inconsistencies found in this study.

However, proceeding from the approximate equation (2) we find:

$$\frac{d\xi}{dt} - \frac{d\xi_0}{dt} = -l\left(\mathrm{div}_p \underline{V} - \mathrm{div}_p \underline{V}_0\right) \qquad \ldots\ldots\ldots\ldots (3)$$

$$\text{but } \frac{d\xi}{dt} - \frac{d\xi_0}{dt} \equiv \frac{\partial \xi'}{\partial t} + \underline{V}.\nabla(\xi + l) - \underline{V}_0.\nabla(\xi_0 + l)$$

$$= \frac{\partial \xi'}{\partial t} + (\underline{V}_0 + \underline{V}').\nabla(\xi_0 + \xi' + l) - \underline{V}_0.\nabla(\xi_0 + l)$$

$$= \frac{\partial \xi'}{\partial t} + \underline{V}_0.\nabla\xi' + \underline{V}'.\nabla\xi_0 + \underline{V}'.\nabla(2\xi_0 + \xi' + l)$$

From equations (1) and (3) we see that the left-hand side of this equation and the last term on the right-hand side are equal, so we are left with:

$$-\frac{\partial \xi'}{\partial t} = \underline{V}_0 . \nabla \xi' - \underline{V}'. \nabla \xi_0$$

or, since the direction of shear is constant with height, following Sutcliffe:

$$-\frac{\partial \xi'}{\partial t} = V_0 \frac{\partial \xi'}{\partial s} - V' \frac{\partial \xi_0}{\partial s} \qquad \ldots\ldots\ldots\ldots (4)$$

This then is a condition of the expression for development in equation (1), namely, that local change of thermal vorticity is due to the difference in advection of thermal vorticity by the surface wind and of surface vorticity by the thermal wind. An alternative which may be implicit in Sutcliffe's analysis is to regard this equation as a residual, negligible in comparison with the terms he takes into account but on the other hand the importance of the term $V' \frac{\partial \xi_0}{\partial s}$ - the "thermal steering effect" - is stressed.

/The

## The Vertical Variation terms.

Sutcliffe (1947) derives the vorticity equation in the form:

$$\frac{d}{dt}(\xi+l) = -(\xi+l)\,\mathrm{div}_p\underline{V} - \frac{\partial v}{\partial p}\left(\frac{\partial}{\partial x}\frac{dp}{dt}\right)_p + \frac{\partial u}{\partial p}\left(\frac{\partial}{\partial y}\frac{dp}{dt}\right)_p$$

whence

$$\frac{d}{dt}(\xi+l) = -(\xi+l)\,\mathrm{div}_p\underline{V} - \nabla_p\frac{dp}{dt}\times\frac{\partial\underline{V}}{\partial p} \qquad (5)$$

Sutcliffe (1947) argues that the sum: $\left(\xi\,\mathrm{div}_p\underline{V} + \nabla_p\frac{dp}{dt}\times\frac{\partial\underline{V}}{\partial p}\right)$ is negligible but Sherman (1952) has shewn that terms equivalent to $\nabla_p\frac{dp}{dt}\times\frac{\partial\underline{V}}{\partial p}$ are of the same order of magnitude as the sum of the advective terms: $\left[\underline{V}.\nabla_p(\xi+l) + \frac{dp}{dt}\frac{\partial}{\partial p}(\xi+l)\right]$ and the term of the form $(\xi+l)\,\mathrm{div}_p\underline{V}$. Reed (1951) made some computations to show that both the terms on the right-hand side of equation (5), or term of that type, were of equal importance. McVittie (1952) also shews that the magnitude of these two terms are comparable and that both should be retained in the scalar vorticity equation.

The attempt was made to evaluate the terms on the right-hand side of equation (5) on two occasions of special interest and the results are given in Table I. Even if the computations proved unreliable as to the exact values of the terms, it was hoped that at least they would give some guide as to their relative order of magnitude. Divergence and vorticity were computed by Bellamy's (1949) "objective" method, the values in Table I being given in units of hour$^{-2}$.

Values found for $\nabla_p\frac{dp}{dt}\times\frac{\partial\underline{V}}{\partial p}$ compare well with those found by Reed (1951) for his equivalent term, namely, $10^{-9}$ to $10^{-8}$ sec$^{-2}$. In the table it may also be seen that the sum of $\xi\,\mathrm{div}_p\underline{V}$ and $\nabla_p\frac{dp}{dt}\times\frac{\partial\underline{V}}{\partial p}$ is not usually zero and that $-l\,\mathrm{div}_p\underline{V}$ is not always the highest absolute magnitude. Eady's (1952) point that errors due to the **approximation** would be larger at higher levels is also confirmed.

When the values of the individual time rate of change of the vorticity at 1000 mb. relative to that at 500 mb. are plotted, the charts obtained are as shewn in Figures 13 and 14. By comparing Figure 13 with Figures 1 and 4 it may be seen that the area of increasing cyclonic vorticity over southern Ireland at 0300 G.M.T. was followed at 0900 G.M.T. by the development of a small secondary in that area, while the area of increasing vorticity over eastern England was associated with the major front of the depression. In Figure 14 the area of increasing cyclonic vorticity off southwest Ireland was followed by the advance into that area of a depression from the Atlantic with its associated front advancing into southwest England. To the extent illustrated by these two examples the computations may be taken as giving real results.

/The

## The term Curl $\left(\frac{\partial \underline{V}}{\partial p}\cdot\frac{dp}{dt}\right)$

Returning to equation (5), we get, by transferring the term $\frac{dp}{dt}\frac{\partial}{\partial p}(\zeta+l)$ from the left to the right-hand side:

$$\left[\frac{d}{dt}(\zeta+l)\right]_p = -(\zeta+l)\,\mathrm{div}_p\underline{V} - \mathrm{curl}_p\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right) \qquad (6)$$

This reduces to:

$$\left[\frac{d}{dt}(\zeta+l)\right]_p = -(\xi+l)\,\mathrm{div}_p\underline{V} \qquad (7)$$

if:

(a) the motion is barotropic, though in this case the term on the right-hand side is small.

(b) the thermal wind is zero, i.e. if the motion is "solid". This is equivalent to (a).

(c) the thermodynamical development represented by $\frac{dp}{dt}$ is zero, i.e. if the vertical motion is zero.

(d) if $\mathrm{Curl}_p\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right) \equiv 0$ neither $\frac{\partial \underline{V}}{\partial p}$ nor $\frac{dp}{dt}$ being zero. In this event $\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}$ must be the gradient of a scalar point function, its direction being determined by that of the thermal wind.. Then also the vorticity equation is similar to that of "solid" motion or of motion without dynamical development. We can say, defining $P \equiv -\mathrm{curl}\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right)$ —

(1) On an isobaric surface where there is no divergence, the sign and magnitude of P will determine those of the individual time rate of change of absolute vorticity.

(2) Generally, the individual time rate of change of absolute vorticity will be modified by P, which, when positive will contribute cyclonic vorticity to the change, and, when negative will contribute anticyclonic vorticity to the change.

The magnitude of the contribution of P to the time rate of change of vorticity on any particular occasion cannot readily be estimated since it is a function of $\frac{dp}{dt}$, perhaps the most important unknown in the equations of motion. However, we may arrive at a qualitative estimate of the effect of P if we consider it in its other form:

$$P = -\mathrm{curl}_p\left(\frac{\partial \underline{V}}{\partial z}w\right) = \frac{\partial \underline{V}}{\partial z}\times\nabla_p w - w\,\mathrm{curl}\frac{\partial \underline{V}}{\partial z} \qquad (8)$$

where w is vertical motion relative to the pressure surface and $\frac{\partial \underline{V}}{\partial z}$ is the thermal wind. For reference we may call the first of the two terms on the right-hand side of the identity (8) $P_1$ and the second, together with its minus sign $P_2$ so that $P = P_1 + P_2$.

Positive and negative values of $w$ will be shewn respectively by areas of continuous rainfall and by areas of subsidence, the former normally being associated at levels near the surface with cyclonic vorticity and the latter with anticyclonic vorticity. A qualitative estimate of P may be made by reference to such areas. The positive direction of the gradient of $w$ will be shown by the perpendiculars to the boundaries of these areas, from the area of no-rain towards the rain and from the area of subsidence towards that of no-subsidence. The thermal wind vector is taken as positive in the direction of the thermal wind. Then a selection of possible cases may be considered:-

(1) $w$ positive - i.e. cyclonic vorticity near the surface.

(a) $P_1$ positive - i.e. air rising at an increasing rate from right to left of the thermal wind. $P_2$ will be positive if the vorticity of the thermal wind is anticyclonic, and the contribution of P to the change in vorticity will be positive (or cyclonic). It is possible to envisage such a series of conditions near a warm front on the flank of a thermal ridge and they may have relevance in the formation of warm front "breakaway" depressions, though this requires to be verified.

(b) $P_1$ negative - i.e. air rising at an increasing rate from left to right of the thermal wind. $P_2$ will be negative if the vorticity of the thermal wind is cyclonic and the contribution of P to the change in vorticity will be negative (or anticyclonic). Such a series of conditions could occur near a cold front on the flank of a thermal trough - with a resulting ridge of high pressure developing behind it.

(2) $w$ negative - i.e. anticyclonic vorticity near the surface.

(a) $P_1$ positive - i.e. air subsiding at an increasing rate from left to right of the thermal wind. $P_2$ will be positive if the vorticity of the thermal wind is cyclonic and the contribution of P to the change in vorticity will be cyclonic. Such a series of conditions could occur on the periphery of an anticyclone underlying an upper thermal ridge, with waves passing round the anticyclone.

(b) $P_1$ negative - i.e. air subsiding at an increasing rate from right to left of the thermal wind. $P_2$ will be negative if the vorticity of the thermal wind is anticyclonic and the contribution of P to the change in vorticity will be anticyclonic. Such a series of conditions could occur with the extension of an anticyclone or ridge of high pressure.

(3) Constant thermal wind, - i.e. $\mathrm{Curl}_p \frac{\partial V}{\partial z} = 0$.

In this case the contribution of P to the change in vorticity will be cyclonic with subsiding air on the warm side of the thermal wind, as on the south side of a west - east airstream, or with rising air on the north side of such a stream. With rising air to the south or subsiding air to the north then the contribution of P would be anticyclonic.

(4) $\mathrm{div}_p \underline{V} = 0$ - i.e. $w$ constant for the pressure thickness throughout which the thermal wind is taken. As indicated above, cyclonic or anticyclonic vorticity would develop in this case according as the sign of P is positive or negative. If $w$ were constant also over an area then, with rising air, existing cyclonic vorticity would be increased in a region of anticyclonic thermal vorticity and diminished in a region of cyclonic thermal vorticity.

/Conclusion

## Conclusion.

It is shown that seeming discrepancies in the application of Sutcliffe's (1947) ideas of development may be due to the employment of an over-simplified scalar vorticity equation and that, in particular, neglect of the vertical variation term may not be justified. This confirms the view of McVittie (1952) who appears sceptical in regard to its usual neglect. A form of this term is derived with no restrictive conditions on the motion of the atmosphere apart from those implicit in the usual two-dimensional frictionless equations of motion: $-\mathrm{Curl}_p\left(\frac{\partial V}{\partial p}\frac{dp}{dt}\right)$ . Defined as P, for convenient reference, it is shewn that its contribution to the time rate of change of vorticity may be profitably considered in forecasting.

A few examples have been given of the possible influence of P and a systematic study of that influence would appear to be justified. In particular the behaviour of depressions and anticyclones relative to the jet stream could perhaps be clarified and also the behaviour of depressions in cold pools and of anticyclones in warm ridges.

As for the day to day use in forecasting of the term P, provided areas of rising and subsiding air could be delineated on the "thickness" charts, qualitative assessments could be made probably with an improvement on present forecasting results.

It may also be that P could be assessed for mean charts over extended periods, with possible gain to the forecaster concerned with such periods.

## Acknowledgments.

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# TABLE I

## 4 FEB 1951 0300 GMT

### 950 MB.

| | 55.5N<br>16.6W | 57.3N<br>12.4W | 55.7N<br>08.0W | 53.3N<br>07.5W | 56.5N<br>05.7W | 54.9N<br>04.4W | 52.8N<br>05.0W | 54.2N<br>02.0W | 52.5N<br>01.5W | 52.6N<br>12.0W | 51.8N<br>03.5W |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $-\xi\,\mathrm{div}\,\underline{V}$ | -0.048 | +0.009 | -0.014 | -0.029 | +0.003 | -0.006 | -0.022 | -0.026 | 0 | +0.028 | +0.021 |
| $-\ell\,\mathrm{div}\,\underline{V}$ | -0.086 | -0.044 | +0.039 | +0.071 | +0.009 | -0.052 | -0.05 | +0.084 | +0.004 | +0.029 | +0.066 |
| $-\overline{\zeta+\ell}\,\mathrm{div}\,\underline{V}$ | -0.134 | -0.035 | +0.025 | +0.042 | +0.012 | -0.058 | -0.072 | +0.058 | +0.004 | +0.057 | +0.087 |
| $-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$ | +0.006 | +0.008 | +0.01 | +0.002 | +0.01 | -0.015 | -0.015 | -0.01 | +0.015 | -0.01 | -0.003 |
| $\frac{d}{dt}(\zeta+\ell)$ | -0.128 | -0.027 | +0.035 | +0.044 | 0 | -0.073 | -0.087 | +0.048 | +0.019 | +0.047 | +0.084 |

### 500 MB.

| | 55.5N<br>16.6W | 57.3N<br>12.4W | 55.7N<br>08.0W | 53.3N<br>07.5W | 56.5N<br>05.7W | 54.9N<br>04.4W | 52.8N<br>05.0W | 54.2N<br>02.0W | 52.5N<br>01.5W | 52.6N<br>12.0W | 51.8N<br>03.5W |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $-\zeta\,\mathrm{div}\,\underline{V}$ | -0.008 | -0.02 | -0.011 | -0.011 | +0.01 | -0.003 | -0.031 | 0 | +0.003 | +0.067 | -0.005 |
| $-\ell\,\mathrm{div}\,\underline{V}$ | -0.013 | -0.057 | +0.069 | +0.025 | +0.013 | -0.009 | -0.042 | 0 | -0.021 | +0.098 | -0.008 |
| $-\overline{\zeta+\ell}\,\mathrm{div}\,\underline{V}$ | -0.021 | -0.077 | +0.058 | +0.014 | +0.023 | -0.012 | -0.073 | 0 | -0.018 | -0.165 | -0.013 |
| $-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$ | +0.07 | +0.05 | +0.03 | -0.02 | +0.08 | -0.05 | 0 | -0.03 | +0.006 | +0.02 | +0.08 |
| $\frac{d}{dt}(\zeta+\ell)$ | +0.051 | -0.027 | +0.088 | -0.006 | +0.103 | -0.062 | -0.073 | 0 | -0.012 | +0.145 | +0.067 |
| $\left(\frac{d\xi}{dt}\right)_{950} - \left(\frac{d\xi}{dt}\right)_{500}$ | -0.179 | 0 | -0.053 | +0.05 | -0.103 | -0.011 | -0.014 | +0.048 | +0.031 | -0.098 | +0.017 |

## 3 OCT 1952 1500 GMT

### 950 MB.

| | 57.3N<br>13.6W | 59.8N<br>07.7W | 55.3N<br>10.9W | 56.5N<br>05.5W | 58.3N<br>03.5W | 53.2N<br>11.8W | 53.4N<br>06.5W | 54.8N<br>04.3W | 56.4N<br>01.0W | 52.0N<br>06.0W | 54.3N<br>01.6W | 51.8N<br>03.5W | 52.5N<br>01.0W |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $-\xi\,\mathrm{div}\,\underline{V}$ | +.002 | -.007 | -.006 | -.003 | -.001 | +.001 | +.001 | -.002 | +.013 | 0 | -.003 | -.004 | -.001 |
| $-\ell\,\mathrm{div}\,\underline{V}$ | -.022 | -.077 | +.022 | +.026 | -.044 | -.004 | -.008 | +.017 | -.039 | +.012 | +.008 | +.008 | +.008 |
| $-\overline{\zeta+\ell}\,\mathrm{div}\,\underline{V}$ | -.02 | -.084 | +.016 | +.023 | -.045 | -.003 | -.007 | +.015 | -.026 | +.012 | +.005 | +.004 | +.007 |
| $-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$ | -.005 | -.005 | +.0009 | -.002 | -.005 | +.003 | -.0008 | +.0003 | +.0004 | -.0018 | -.0001 | +.0011 | -.0006 |
| $\frac{d}{dt}(\zeta+\ell)$ | -.025 | -.089 | +.0169 | +.021 | -.05 | 0 | -.0078 | +.0153 | -.0256 | +.0102 | +.0049 | +.0051 | +.0064 |

### 500 MB.

| | 57.3N<br>13.6W | 59.8N<br>07.7W | 55.3N<br>10.9W | 56.5N<br>05.5W | 58.3N<br>03.5W | 53.2N<br>11.8W | 53.4N<br>06.5W | 54.8N<br>04.3W | 56.4N<br>01.0W | 52.0N<br>06.0W | 54.3N<br>01.6W | 51.8N<br>03.5W | 52.5N<br>01.0W |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $-\xi\,\mathrm{div}\,\underline{V}$ | +.003 | +.031 | -.003 | +.01 | -.001 | 0 | +.001 | 0 | -.001 | -.002 | +.007 | +.006 | -.005 |
| $-\ell\,\mathrm{div}\,\underline{V}$ | -.057 | +.063 | +.009 | -.026 | -.031 | 0 | -.004 | 0 | -.004 | +.12 | +.017 | -.033 | -.053 |
| $-\overline{\zeta+\ell}\,\mathrm{div}\,\underline{V}$ | -.054 | +.094 | +.006 | -.016 | -.032 | 0 | -.003 | 0 | -.005 | +.01 | +.024 | -.027 | -.058 |
| $-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$ | +.03 | -.03 | -.02 | +.01 | 0 | -.05 | +.02 | +.004 | -.03 | -.03 | +.001 | -.04 | +.01 |
| $\frac{d}{dt}(\zeta+\ell)$ | -.024 | +.064 | -.014 | -.006 | -.032 | -.05 | +.017 | +.004 | -.035 | -.02 | +.025 | -.067 | -.048 |
| $\left(\frac{d\xi}{dt}\right)_{950} - \left(\frac{d\xi}{dt}\right)_{500}$ | -.001 | -.153 | +.031 | +.027 | -.018 | +.05 | -.025 | +.011 | +.009 | +.03 | -.02 | +.072 | +.054 |

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M.R.P. 829.

FIG. 1.

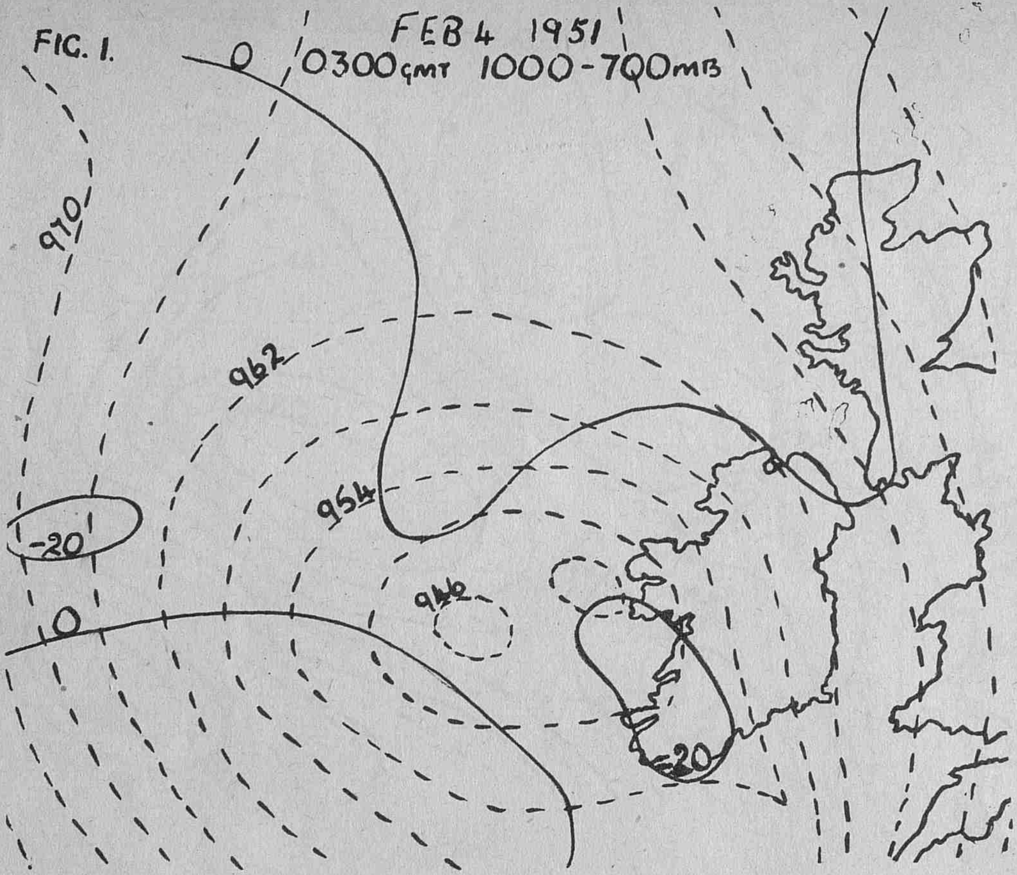

FEB 4 1951
0300GMT 1000-700MB
0
970
962
954
946
-20
0
-20

FIG. 2

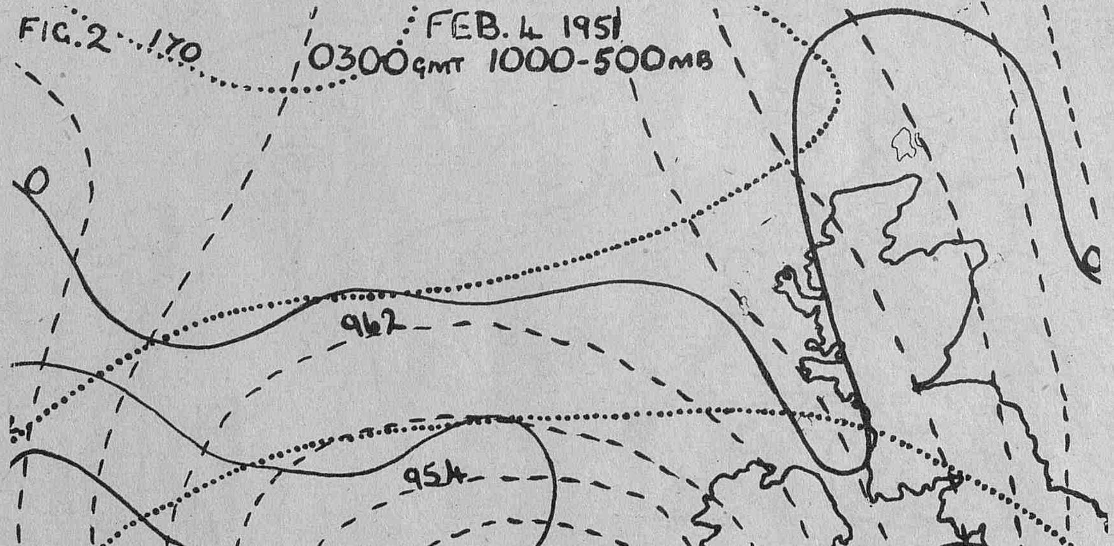

170
FEB. 4 1951
0300GMT 1000-500MB
0
962
954
946
-40
-20
0
+20
+40
-20
178
0

FIG. 3. FEB 4 1951
0300 GMT 1000-300 MB

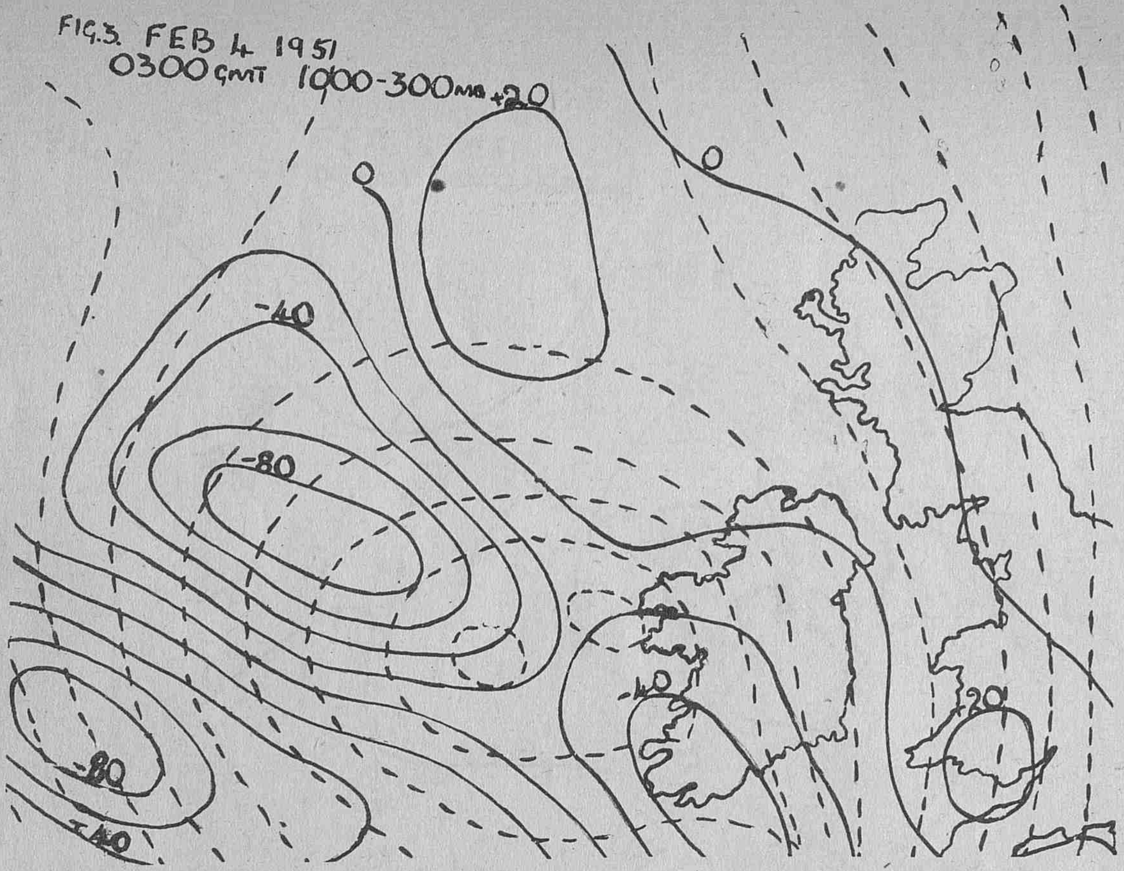
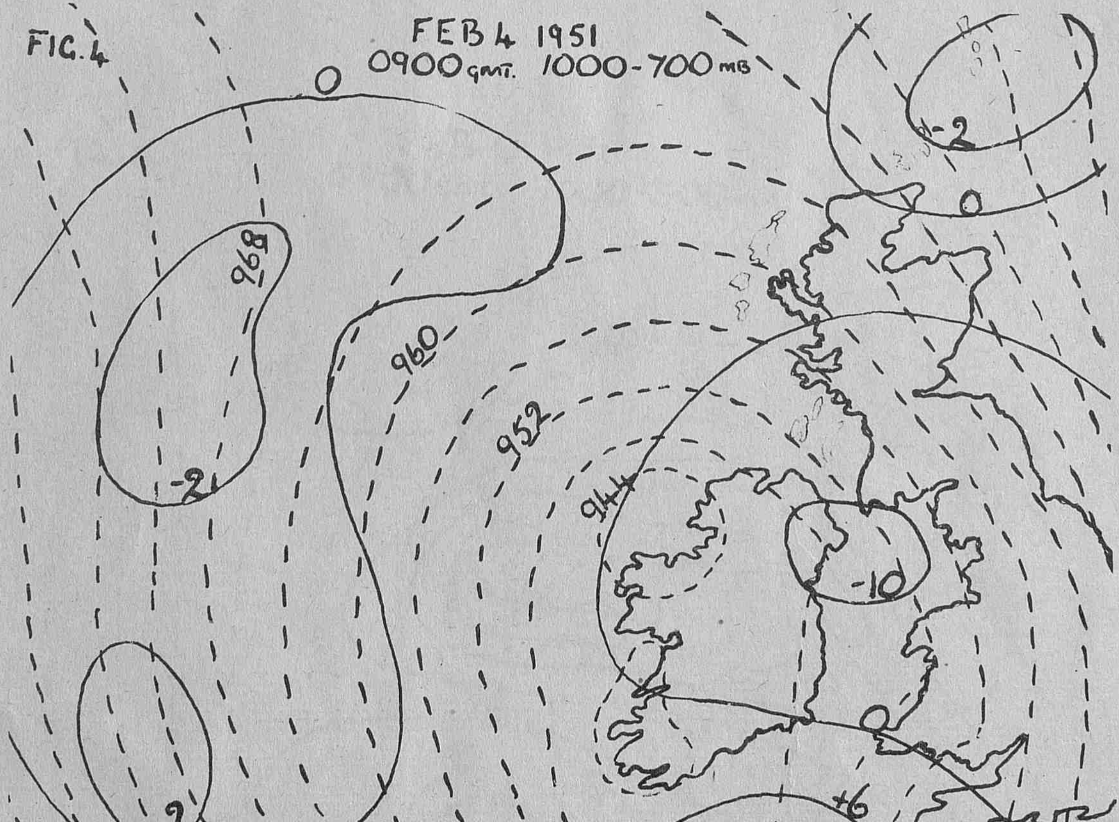
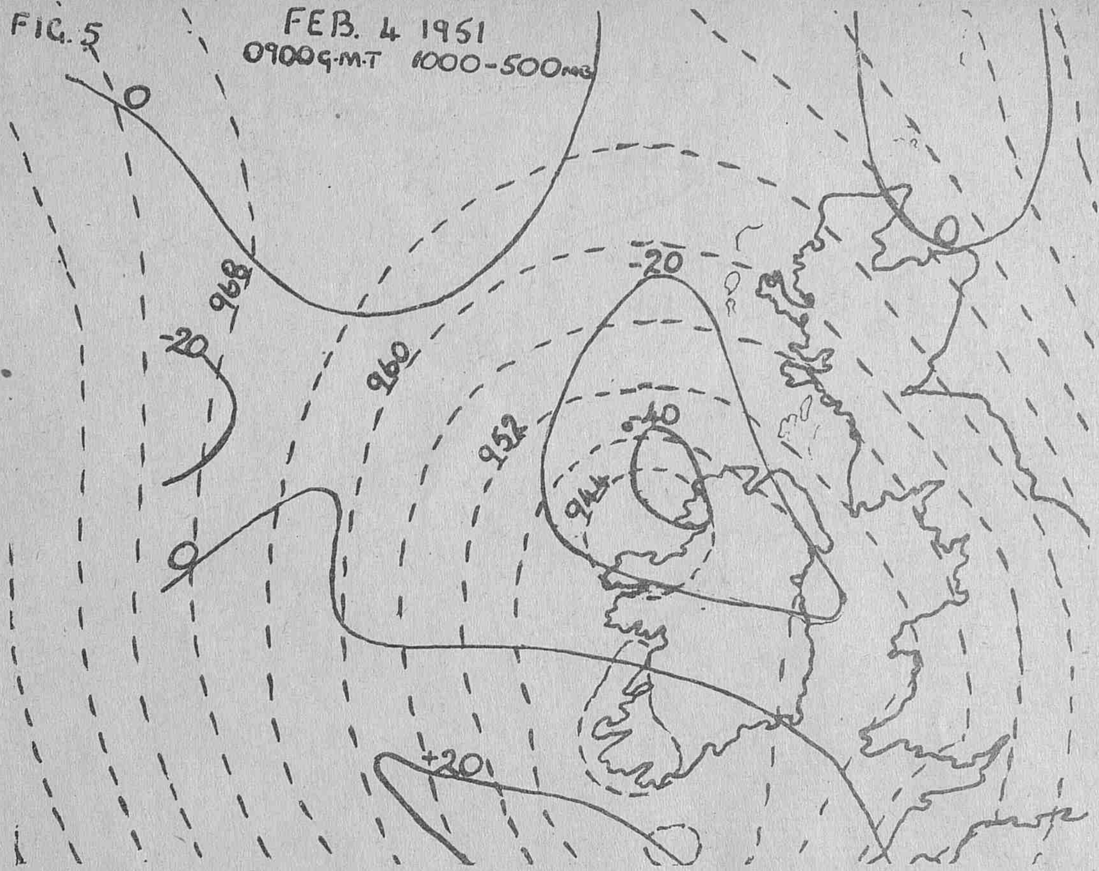

+20
0
0
-40
-80
-80
-40
-40
-40
20

FIG. 4 FEB 4 1951
0900 GMT. 1000-700 MB


0
968
960
952
944
-2
-2
-2
0
-10
0
+6

M.R.P. 829.

FIG. 5


FEB. 4 1951
0900 G.M.T 1000-500 MB
0
968
-20
960
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952
-40
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FIG. 6.

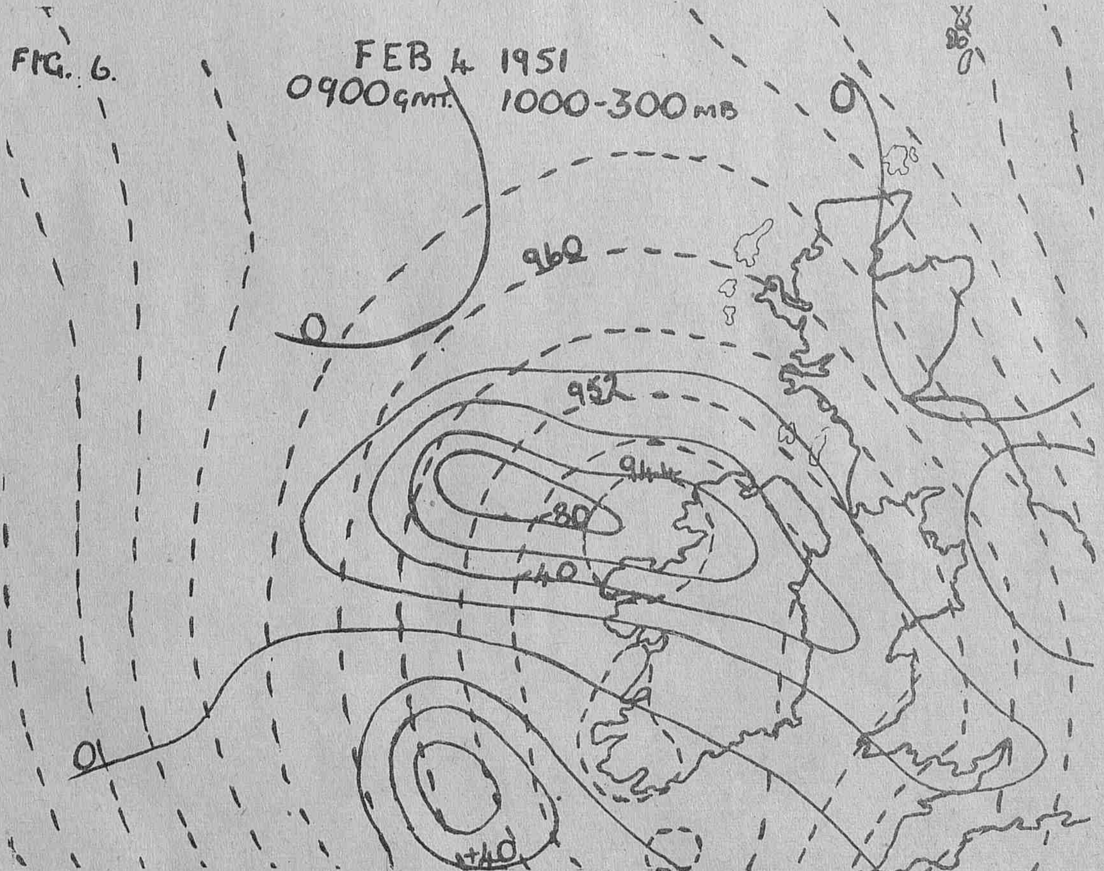

FEB 4 1951
0900 GMT. 1000-300 MB
0
0
960
0
952
944
-80
-40
0
+40

FIG. 7. 1500 GMT FEB 4 1951 1000-700 MB

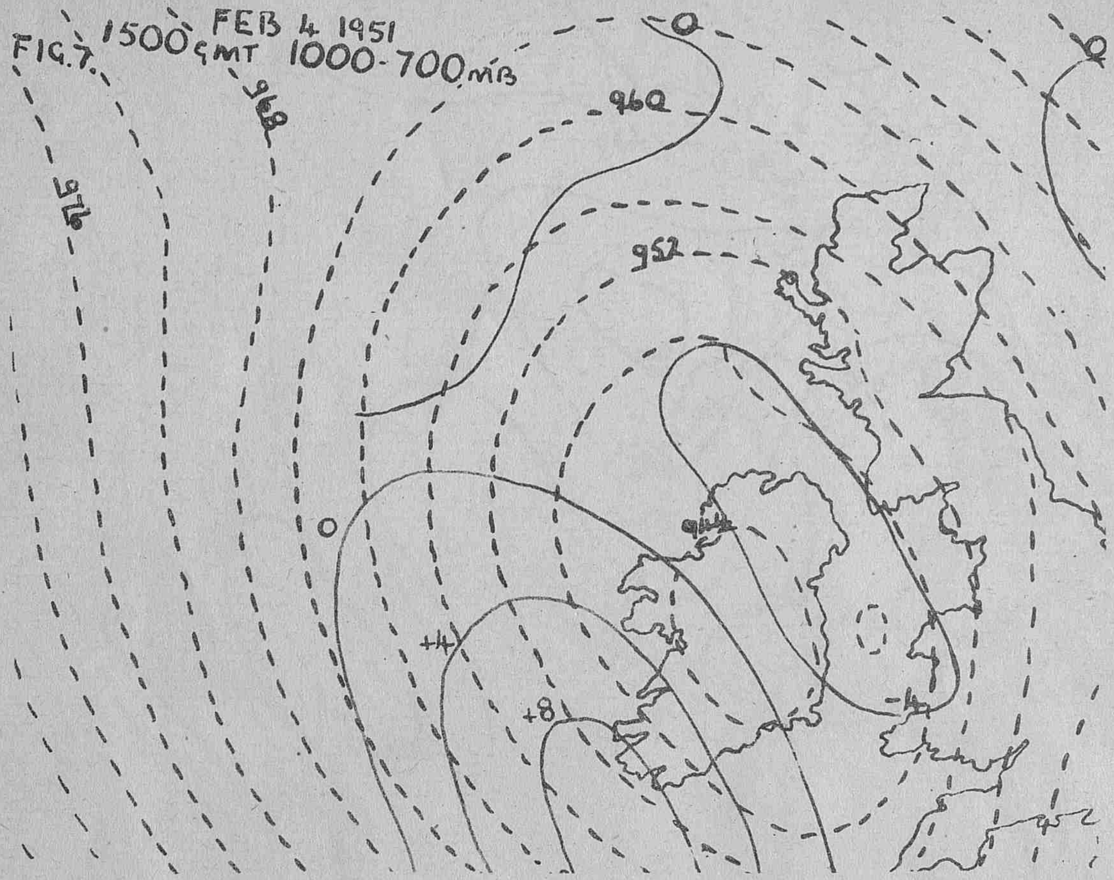
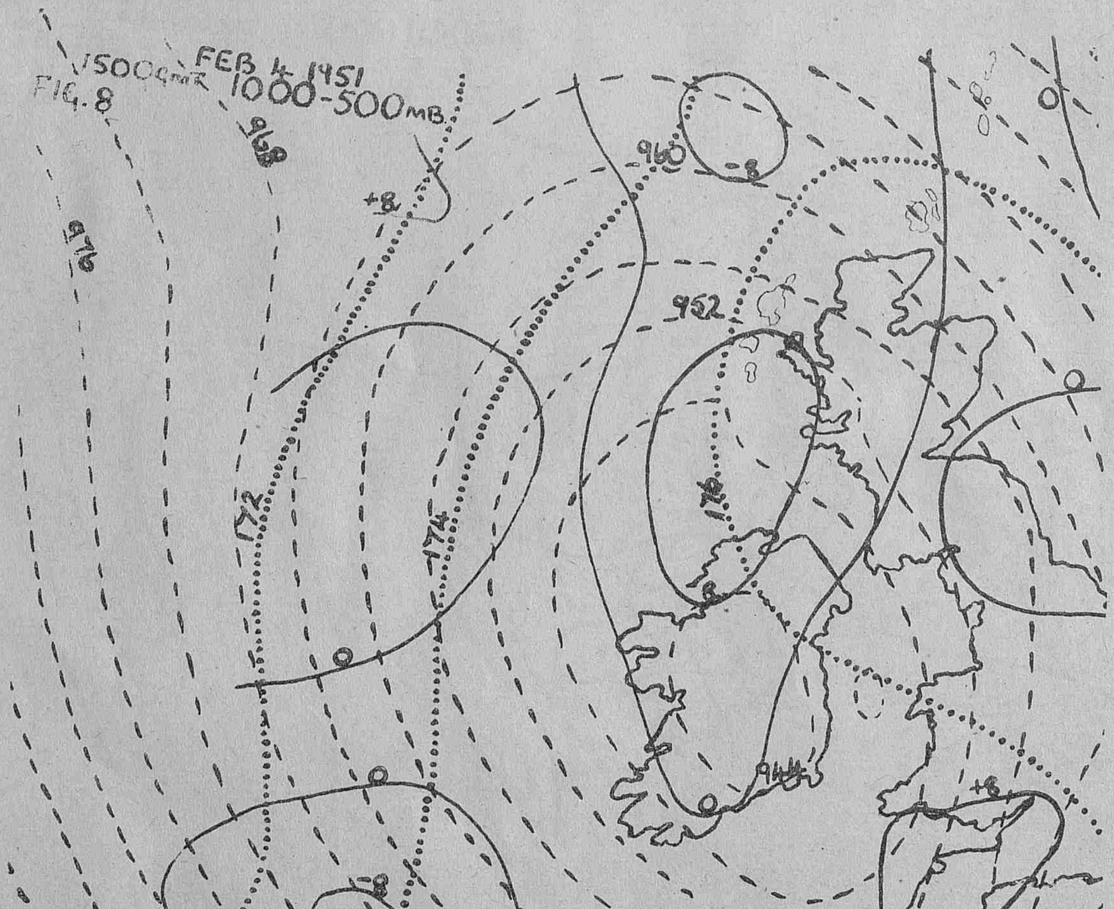

976
968
960
952
944
0
0
0
+4
+8
-4
952

FIG. 8 1500 GMT FEB 4 1951 1000-500 MB.


976
968
960
952
944
172
174
176
+8
-8
+8
-8
0
0
0
0
0

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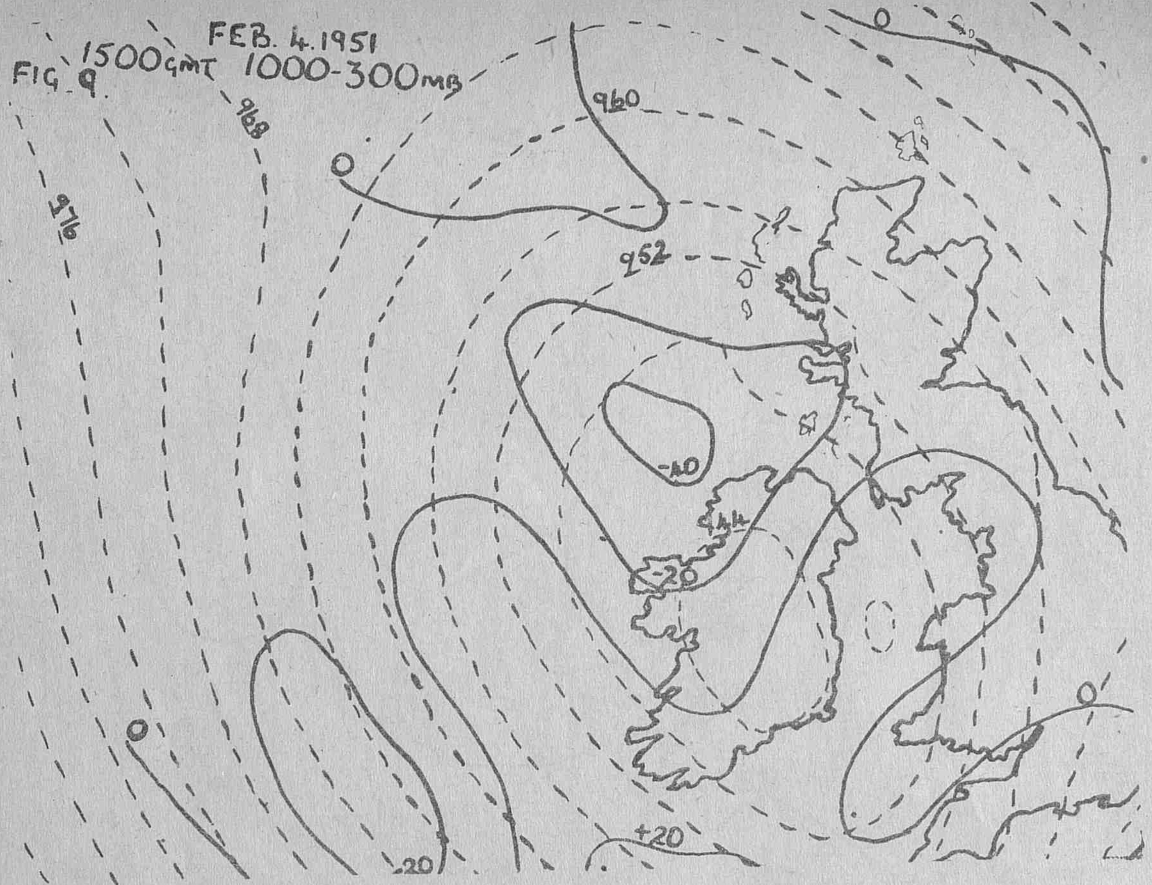
FIG. 9. 1500 GMT FEB. 4. 1951 1000-300 MB

960
968
976
952
944
-40
-20
-20
+20
0
0
0
0

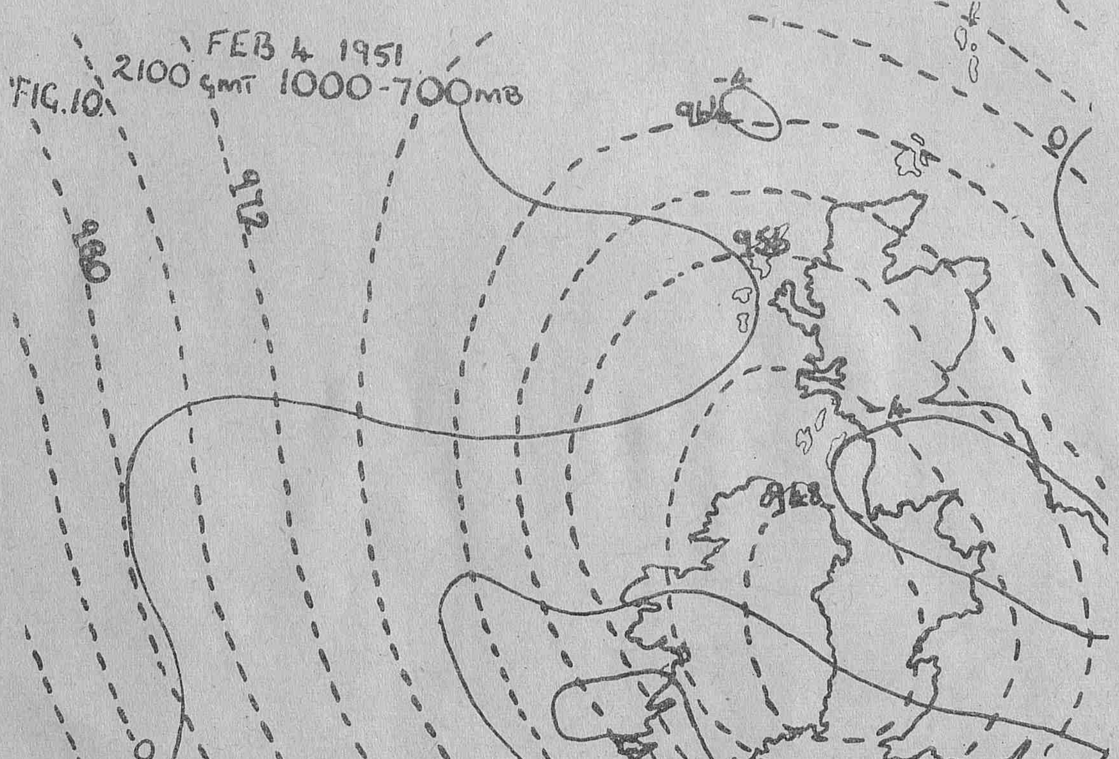
FIG. 10. 2100 GMT FEB 4 1951 1000-700 MB

980
972
964
956
948
-4
-4
+4
0
0

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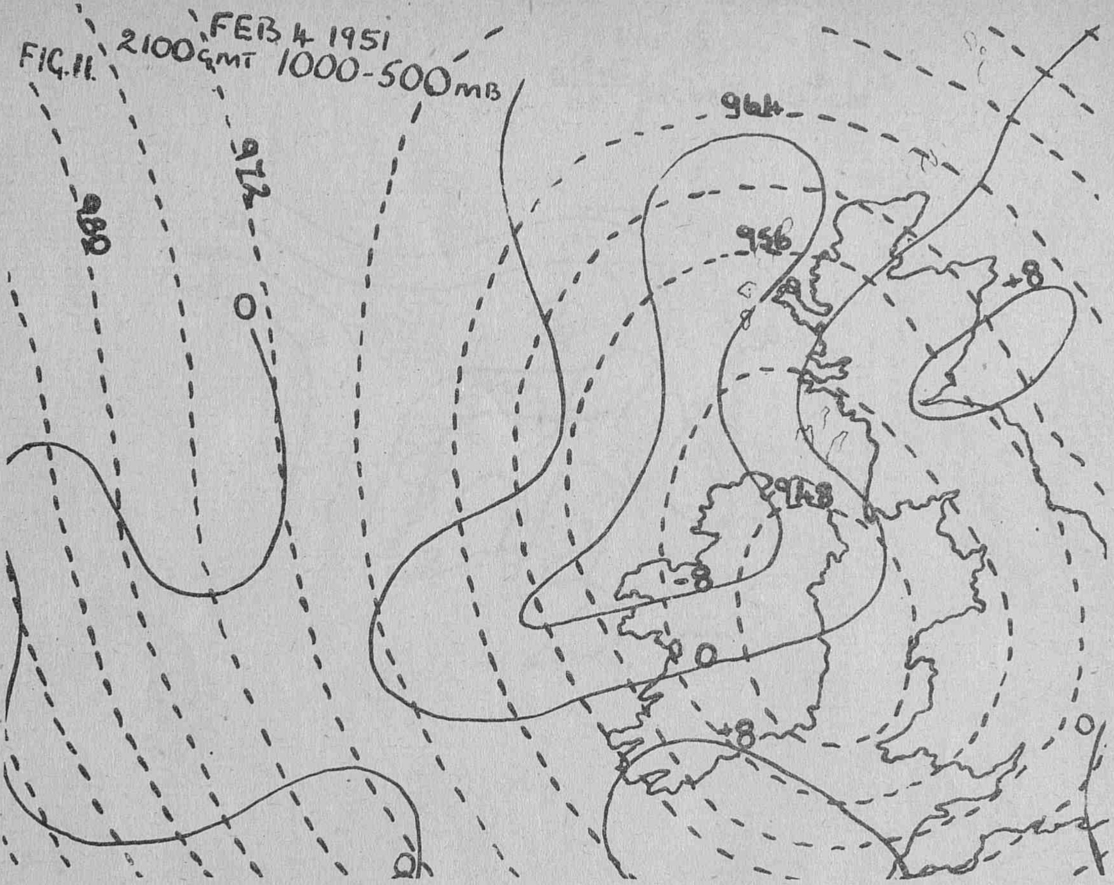
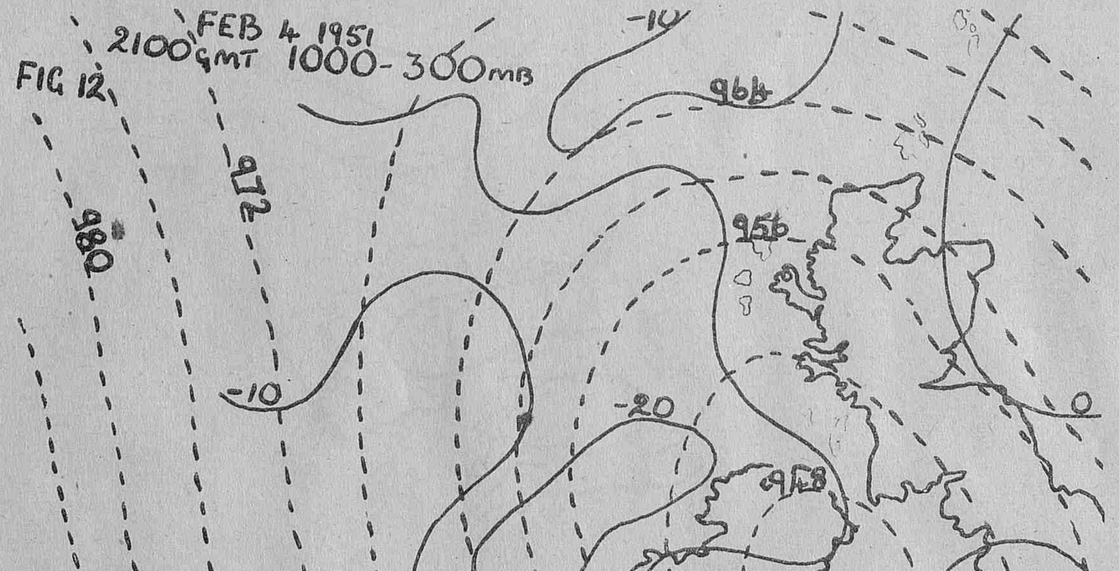
FIG. 11.



2100 GMT
FEB 4 1951
1000-500 mB
980
972
964
956
948
0
0
0
0
0
-8
+8
+8

FIG 12

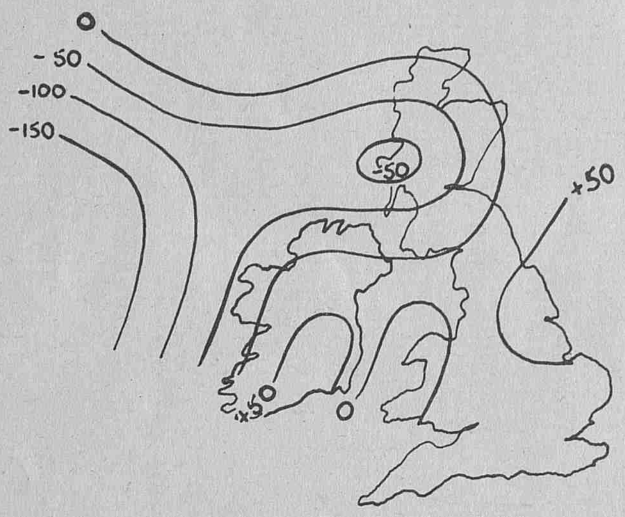


2100 GMT
FEB 4 1951
1000-300 mB
980
972
964
956
948
-10
-10
-10
-20
0
0
+10
+10
+20

FEB 4 1951 0300 GMT

FIG 13.

$\frac{d(\zeta+l)}{dt}$ in units $10^{-3}$ $hr^{-2}$

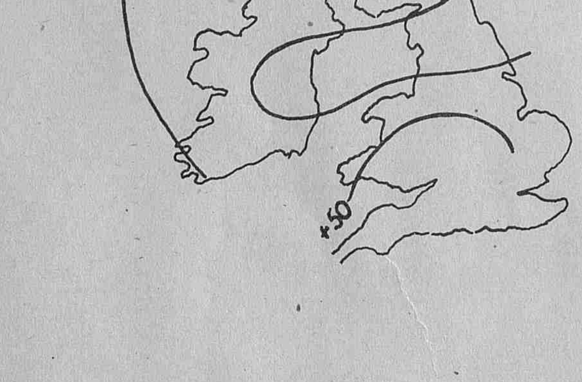


0
-50
-100
-150
-50
+50
0
+50
0

OCT 3 1952 1500 GMT

FIG 14

$\frac{d(\zeta+l)}{dt}$ in units $10^{-3}$ $hr^{-2}$



-150
-100
-50
0
+50
+50

M.O.15/R/7

[illegible]

Secretary,
Meteorological Research Committee

Meteorological Office (M.O.15),
Air Ministry,
London, W.C.2.

22nd February, 1957.

G.5280/PG/2/57/170

[illegible]

To all recipients of M.R.P. 830 (AD 39446)

Will you please attach the enclosed addendum to your copy of M.R.P. 830 and also note that this paper has now been declassified.

H W L Absalom
Secretary,
Meteorological Research Committee

Meteorological Office (M.O.15),
Air Ministry,
London, W.C.2.

22nd February, 1957.

G.5280/PG/2/57/170

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Thus if we take axes moving with the aircraft, with the origin at the points where the droplets are ejected from the wake, then with x measured downwind, y across wind and z vertically upward the equation which states that the advection is balanced by lateral and vertical diffusion is as in 2.04

$$U\frac{\partial\chi}{\partial x} = \frac{\partial}{\partial y}\left(K_y\frac{\partial\chi}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z\frac{\partial\chi}{\partial z}\right) \tag{6.01}$$

where U is the velocity of the aircraft relative to the air.

Now in the case in which diffusion takes place immediately behind the aircraft the expressions used for the coefficients of eddy diffusion in the y and z directions respectively were

$$K_y = \frac{UC^2|y|^{1-m}}{(1+m)^2} \tag{6.02}$$

$$\text{and} \quad K_z = \frac{UC^2|z|^{1-m}}{(1+m)^2} \tag{6.03}$$

and the solution of equation of eddy diffusion

$$U\frac{\partial\chi}{\partial x} = \frac{\partial}{\partial y}\left\{\frac{UC^2|y|^{1-m}}{(1+m)^2}\frac{\partial\chi}{\partial y}\right\}+\frac{\partial}{\partial z}\left\{\frac{UC^2|z|^{1-m}}{(1+m)^2}\frac{\partial\chi}{\partial z}\right\} \tag{6.04}$$

with the appropriate boundary conditions, was used to discuss the problem.

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## Addendum to M.R.P. 830

## The prediction of contrails from jet aircraft

In para. 5(c) of M.R.P. 830 it was noted that for trail lengths in excess of about 250 metres the predicted lengths were less than those observed and it was suggested that this was a consequence of the assumption that the trail from an aircraft moving through the air with a velocity of U cm sec$^{-1}$ is the same as that from a fixed point in the atmosphere in which the mean speed is U cm sec$^{-1}$. It was pointed out that this assumption could only be true for a limited distance downwind from the aircraft for in the immediate turbulent wake it was found that the diffusion coefficients were the same as those in the free air apart from the fact that in the former U the velocity of the aircraft relative to the air replaced V the mean wind speed in the latter, and it was suggested that the turbulent wake extended for about 250 metres.

With a jet aircraft in which trail lengths are of the order of a mile or more and where a knowledge of the outline of the contrail in the vicinity of the exhaust is of little significance, the problem of predicting the occurrence, length, width and persistence of contrails can be dealt with very simply if it is assumed that the mixing takes place in the turbulent wake immediately behind the aircraft and that the droplets resulting from this mixing are ejected from the end of this wake into the free atmosphere where they diffuse normally.

Thus if we take axes moving with the aircraft, with the origin at the points where the droplets are ejected from the wake, then with x measured downwind, y across wind and z vertically upward the equation which states that the advection is balanced by lateral and vertical diffusion is as in 2.04

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left(K_y \frac{\partial \chi}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial \chi}{\partial z}\right) \qquad 6.01$$

where U is the velocity of the aircraft relative to the air.

Now in the case in which diffusion takes place immediately behind the aircraft the expressions used for the coefficients of eddy diffusion in the y and z directions respectively were

$$K_y = \frac{UC^2|y|^{1-m}}{(1+m)^2} \qquad 6.02$$

and $$K_z = \frac{UC^2|z|^{1-m}}{(1+m)^2} \qquad 6.03$$

and the solution of equation of eddy diffusion

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left\{\frac{UC^2|y|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial y}\right\} + \frac{\partial}{\partial z}\left\{\frac{UC^2|z|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial z}\right\} \qquad 6.04$$

with the appropriate boundary conditions, was used to discuss the problem.

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In the present case using coefficients of eddy diffusion appropriate to the free air, i.e., with V replacing U, the equation to be solved is

$$U\frac{\partial\chi}{\partial x} = \frac{\partial}{\partial y}\left\{\frac{VC^2|y|^{1-m}}{(1+m)^2}\frac{\partial\chi}{\partial y}\right\} + \frac{\partial}{\partial z}\left\{\frac{VC^2|z|^{1-m}}{(1+m)^2}\frac{\partial\chi}{\partial z}\right\} \qquad 6.05$$

with the same boundary conditions as before.

If however we transform this equation by introducing a new coefficient $C_1$ where

$$UC_1^2 = VC^2 \qquad 6.06$$

then equations 6.05 and 6.04 are the same with $C_1$ replacing C and hence any solutions derived and used in M.R.P. 830 can be used in the present discussion providing $C^2$ wherever it occurs is now replaced by $\frac{VC^2}{U}$ . For example the visible outline of a contrail immediately behind a piston type aircraft is from 3.05,

$$\frac{|z|^{1+m}}{x} = \frac{C^2}{1+m}\log_e\frac{L}{x} \qquad 6.07$$

so that the visible outline of a contrail at some distance behind a jet aircraft is given by

$$\frac{|z|^{1+m}}{x} = \frac{VC^2}{U(1+m)}\log_e\frac{L}{x} \qquad 6.08$$

and again the trail width is a maximum at about 3/7 of its length.

Similarly 3.13 which gives the relationship between the trail length L, drop radius r, fuel consumption per cm of path $\frac{\Phi}{U}$ and amount of water which is condensed as the result of combustion of each gram of fuel $\lambda$ , shows that for the piston type aircraft,

$$L = \frac{\text{const.}}{C^2}\left(\frac{\phi\lambda}{Ur}\right)^{1+m} \qquad 6.09$$

so that in the case of jet aircraft the relationship becomes

$$L = \frac{\text{const.}\,U}{C^2V}\left(\frac{\phi\lambda}{Ur}\right)^{1+m} \qquad 6.10$$

According to details given by Bannon (18) the fuel consumption per cm of path at 30,000 feet for a Meteor IV is approximately the same as that of the Mosquito, details of whose contrails are given in Table IV.

If therefore $\lambda$ is the same in both cases then with a speed of 500 kt a Meteor IV would have produced a visible trail of approximately 2,000 metres, if the wind speed had been 50 kt, as opposed to the visible trail of approximately 200 metres produced by the Mosquito.

Reference

(18) J.K. Bannon 1949 Meteorological Research Papers, Air Ministry, London, No. 489.

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time been proposed for the point source problem are briefly reviewed.

Recent observations of contrails made by the Meteorological Research Flight, Farnborough, in conjunction with the Central Photographic Establishment, Benson, are discussed by means of the most promising of these solutions and the results appear sufficiently encouraging to warrant further study.

## The point source solutions

2. Roberts (2) on the basis of the classical theory of turbulence obtained the following solution for the continuous point source problem

$$\chi = \frac{Q}{4\pi K(x^2+y^2+z^2)^{1/2}} \exp - \frac{(y^2+z^2)}{(x^2+y^2+z^2)^{1/2}} \frac{u}{4K} \qquad 2.00$$

where the origin of the cartesian coordinate system is taken at the point source with x measured down wind, y across wind and z vertically upwards, $\chi$ is the concentration of smoke at x, y, z, K is the (constant) coefficient of eddy diffusion, u is the mean wind speed (assumed invariant with height) and $Q$ is the rate of emission.

Whilst however first tests of predictions of 2.00 against observations of smoke trails shewed that $\chi$ varied directly as $Q$ and inversely as u (which required K to be proportional to the mean wind speed), later observations quoted by Sutton (3) shewed that the peak concentration at y=z=o fell off as $x^{-1.76}$ under adiabatic conditions, whereas 2.00 required that the peak concentration should fall off as $x^{-1}$.

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radius of a contrail at some distance behind a jet aircraft is given by

$$\frac{|z|^{1+m}}{x} = \frac{VC^2}{U(1+m)} \log_e \frac{L}{x} \qquad 6.08$$

and again the trail width is a maximum at about 3/7 of its length.

Similarly 3.13 which gives the relationship between the trail length L, drop radius r, fuel consumption per cm of path $\frac{\Phi}{U}$ and amount of water which is condensed as the result of combustion of each gram of fuel $\lambda$, shows that for the piston type aircraft,

$$L = \frac{\text{const.}}{C^2}\left(\frac{\phi\lambda}{Ur}\right)^{1+m} \qquad 6.09$$

so that in the case of jet aircraft the relationship becomes

$$L = \frac{\text{const.}\ U}{C^2 V}\left(\frac{\phi\lambda}{Ur}\right)^{1+m} \qquad 6.10$$

According to details given by Bannon (18) the fuel consumption per cm of path at 30,000 feet for a Meteor IV is approximately the same as that of the Mosquito, details of whose contrails are given in Table IV.

/If therefore

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S.C.III/161
Revised December, 1953

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Classification Changed
to Unclassified
by authority of
Commandir
Air Weather Service
AWS
By Meteorological Office (no 15) Air Ministry, London
Date 22 Feb 1957

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# The prediction of contrails as a problem of diffusion

by

R. Frost (Meteorological Office)

## Introduction

1. The problem of the formation of contrails which has been the subject of numerous investigations (1) is becoming of increasing importance as the speed and heights at which military aircraft operate increase, but while it is generally accepted that such trails are due to the condensation of water vapour from the engine exhaust no method has been developed which will enable the length, nature and persistence of contrails to be predicted from meteorological observations.

The resemblance between photographs of contrails and photographs of smoke trails from factory chimneys suggested that the solution of the continuous point source problem which had proved fruitful in discussions of the latter problem might afford a possible line of attack upon the former and more difficult problem, and in the present paper the various solutions which have from time to time been proposed for the point source problem are briefly reviewed.

Recent observations of contrails made by the Meteorological Research Flight, Farnborough, in conjunction with the Central Photographic Establishment, Benson, are discussed by means of the most promising of these solutions and the results appear sufficiently encouraging to warrant further study.

## The point source solutions

2. Roberts (2) on the basis of the classical theory of turbulence obtained the following solution for the continuous point source problem

$$\chi = \frac{Q}{4\pi K(x^2+y^2+z^2)^{1/2}} \exp - \frac{(y^2+z^2)}{(x^2+y^2+z^2)^{1/2}} \frac{u}{4K} \qquad 2.00$$

where the origin of the cartesian coordinate system is taken at the point source with x measured down wind, y across wind and z vertically upwards, $\chi$ is the concentration of smoke at x, y, z, K is the (constant) coefficient of eddy diffusion, u is the mean wind speed (assumed invariant with height) and $Q$ is the rate of emission.

Whilst however first tests of predictions of 2.00 against observations of smoke trails shewed that $\chi$ varied directly as $Q$ and inversely as u (which required K to be proportional to the mean wind speed), later observations quoted by Sutton (3) shewed that the peak concentration at y=z=o fell off as $x^{-1.76}$ under adiabatic conditions, whereas 2.00 required that the peak concentration should fall off as $x^{-1}$.

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The failure of the K = constant concept to give a satisfactory approach to the point source problem was attributed to the non homogeneity of the atmospheric eddy structure, but attempts to modify the classical theory in order to take account of the effect of eddies of many sizes acting together met with little success.

O.G. Sutton (4) discarded the K approach and from an extension of G.I. Taylor's (5) statistical theory of diffusion by continuous movements derived the following approximate solution for the continuous point source problem,

$$\chi = \frac{Q}{\pi C_y C_z x^{2-n}} \exp - \left\{ \frac{1}{x^{2-n}} \left( \frac{y^2}{C_y^2} + \frac{z^2}{C_z^2} \right) \right\} \qquad 2{\cdot}01$$

where $C_y$ and $C_z$ were generalised coefficients of diffusion across wind and in the vertical respectively and n was an index of the degree of turbulence, but later (6) suggested that

$$\chi = \frac{Q}{4\left[\Gamma\left(\frac{4-n}{2}\right)\right]^2 C_y^{2-n} C_z^{2-n} u x^{2-n}} \exp - \left\{ \frac{1}{x} \left( \frac{y^{\frac{2}{2-n}}}{C_y^2} + \frac{z^{\frac{2}{2-n}}}{C_z^2} \right) \right\} \qquad 2{\cdot}02$$

gave a better approximation to the observed facts.

Unlike Roberts' solution 2.00, which being derived from a second order differential equation with the appropriate boundary conditions is therefore a unique solution, 2.01 and 2.02 are only two out of an infinite number of possible solutions which can be obtained by the statistical approach. For example if the choice of solution is restricted to a single expression based on the exponential function, Sutton's statistical approach yields

$$\chi = \frac{Q}{4\Gamma(1+\frac{1}{r})\Gamma(1+\frac{1}{s}) C_y^{2/r} C_z^{2/s} u x^{p/r+p/s}} \exp - \left\{ \frac{1}{x^p} \left( \frac{y^r}{C_y^2} + \frac{z^s}{C_z^2} \right) \right\} \qquad 2{\cdot}03$$

an expression which contains five arbitrary constants p, r, s, $C_y$ and $C_z$ and includes 2.01 and 2.02 as particular solutions.

The reasons for selecting any particular solution of 2.03 as the appropriate solution for the point source problem, and the implications of any particular solution which may be selected, are not clear from the statistical approach which is not concerned with the mechanism of eddy motion. When however 2.03 is derived by means of an extension of the classical theory it is possible to see what assumptions are required and by a combination of theory and observation to decide which solution is likely to give the best representation of the observed facts.

Thus, according to the classical theory, the differential equation which expresses the fact that, when a steady state has been reached, horizontal advection is balanced by lateral and vertical diffusion is

$$u\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left(K_y \frac{\partial \chi}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial \chi}{\partial z}\right) \qquad 2{\cdot}04$$

where $K_y$ and $K_z$ are the coefficients of eddy diffusion in the y and z directions respectively, and the solution of this equation which fits the boundary conditions appropriate to the point source problem is given by 2.03, provided that

$$K_y = uC_y^2 \frac{p}{r^2} x^{p-1} |y|^{2-r} \qquad 2{\cdot}05$$

and

$$K_z = uC_z^2 \frac{p}{s^2} x^{p-1} |z|^{2-s} \qquad 2{\cdot}06$$

Hence 2.01 which is derived from 2.03 by writing p=2-n and r=2 is the solution of 2.04 with

$$K_y = uC_y^2 \left(\frac{2-n}{4}\right) x^{1-n} \qquad 2{\cdot}07$$

$$K_z = uC_z^2 \left(\frac{2-n}{4}\right) x^{1-n} \qquad 2{\cdot}08$$

Coefficients of eddy diffusion in the atmosphere which increase with distance downwind from the point at which the smoke is being emitted, are however physically unattractive, since they appear to imply that the atmosphere is aware of the point at which the smoke is being emitted and is able to adjust its eddies accordingly. This consideration would therefore appear to render unacceptable any solution of the type 2.03 with the possible exception of those for which p = 1 (which includes 2.02 as a special case).

With p = 1, 2.03 becomes

$$\chi = \frac{Q}{4\Gamma(1+\frac{1}{r})\Gamma(1+\frac{1}{s}) C_y^{2/r} C_z^{2/s} u x^{1/r + 1/s}} \exp\left\{\frac{1}{x}\left(\frac{y^r}{C_y^2} + \frac{z^s}{C_z^2}\right)\right\} \qquad 2{\cdot}09$$

whilst

$$K_y = \frac{uC_y^2}{r^2} |y|^{2-r} \qquad 2{\cdot}10$$

and

$$K_z = \frac{uC_z^2}{s^2} |z|^{2-s} \qquad 2{\cdot}11$$

so that in this case the coefficients of eddy diffusion are independent of the distance downwind from the point at which the smoke is being emitted, but are dependent upon the distance from the axis.

This at first sight seems nearly as unattractive as the former concept, but as a smoke trail may be regarded as an aggregate of a very large number of smoke puffs of various ages, then, focusing attention on one particular puff as it spreads downwind and taking axes moving with the mean wind speed, with the origin of coordinates at the centre of gravity of the puff, the functional dependence of $K_y$ and $K_z$ upon y and z respectively may be taken as indicating that as the puff moves downwind and expands it progressively comes under the influence of larger and larger eddies, but that only eddies which are comparable

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with or are smaller than the smoke puff at any instant have any effect upon its rate of expansion.

A functional form for $K_z$ similar to 2.11 but with $u_{10}$ at a standard height of 10 m. replacing u was derived by the present writer and used to discuss the problems of diffusion of matter (7) from plane sources situated at the earth's surface, and observations near the ground showed that under adiabatic conditions good agreement was reached with a value of $s = 8/7$. If it is permissible to ignore the variation of wind with height near the surface in order to identify 2.09 with observations of smoke trails from point sources near the ground which shew that $\chi \propto x^{-1.76}$, Sutton (3), it follows that with $s = 8/7$ then, to a high degree of approximation, r also $= 8/7$, so that 2.09 reduces to Sutton's solution 2.02 with $n = \frac{1}{4}$.

Since the variation of wind with height is very marked near the ground only the order of magnitude of $C_y$ or $C_z$ can be obtained by comparing 2.02 with observations of smoke trails near the surface and this gives the order of magnitude as 0.30 $cm^{-\frac{1}{4}}$.

In order to prevent possible confusion between generalised diffusion coefficients which in equations 2.01 and 2.02 are of very different dimensions being $(\text{length})^{\frac{1}{8}}$ in the former and $(\text{length})^{1/14}$ in the latter, $c_y$ and $c_z$ will be used for the former and $C_y$ and $C_z$ for the latter. When c.g.s. units are used however the numerical values of the coefficients are approximately the same.

If $\sigma_y$ is the standard deviation of the smoke particles from the mean at right angles to the wind, then from 2.01

$$\sigma_y^2 = \tfrac{1}{2} c_y^2 x^{2-n} \qquad 2.12$$

whilst from 2.02

$$\sigma_y^2 = \frac{\Gamma\{\frac{3}{2}(2-n)\}}{\Gamma\{\frac{1}{2}(2-n)\}} \left(C_y^2 x\right)^{2-n} \qquad 2.13$$

Sutton (4) who tested out 2.12 against Richardson's (8) data on pilot balloons over distances of several hundred kilometres obtained a value of $n = \frac{1}{4}$ in good agreement with the values of n found from observations of smoke trails over short distances near the ground.

From an analysis of the growth of clouds formed by the bursting of anti-aircraft shells between 1 and 5 km., Sutton (4) on the assumption that $c_x = c_y = c_z = c$ also found that $n = \frac{1}{4}$ and that the c's varied between 0.088 and 0.022 $cm.^{\frac{1}{8}}$, the values of c apparently decreasing with height according to a law which yielded $c = 0.30$ $cm.^{\frac{1}{8}}$ at a height of 1 m. above the surface and $c = 0$ at a height of 10,000 m. above the surface.

The value of $n = \frac{1}{4}$, but not a law of decrease of c with height, was also found by Sartor and others (9) to fit observations of the growth of smoke puffs between 8 m. and 45 m. R.F. Jones (10) who took observations of smoke puffs at heights varying from 2700 m. to 11700 m. and examined the motion of the smoke puffs at intervals of 20 seconds by means of 2.12 obtained values of n which shewed considerable scatter with however a mean value of .15. The values of $\sigma_x$ which he obtained were not normally the same as the values of $\sigma_y$, but the differences were not systematic and the values of $c_x$ and $c_y$ calculated from these values of $\sigma_x$ and $\sigma_y$ shewed no variation with height, wind shear or lapse rate. By plotting all values of log $c_x$ and log $c_y$ combined against values of 2-n , he found that the relationship between c and n was given by

$$c = \frac{10^{3.47}}{10^{2.39(2-n)}} \qquad 2.14$$

whence

$$\sigma^2 = 4.4 \times 10^6 \left(\frac{x}{6 \times 10^4}\right)^{2-n} \qquad 2.15$$

With $n = \frac{1}{4}$, 2.14 gives $c = c_x = c_y = 0.195 \text{ cm}^{\frac{1}{8}}$.
Jones' observations when examined by 2.13 yield $C = C_x = C_y = 0.287 \text{ cm}^{\frac{1}{4}}$

The evidence thus available suggests that diffusion from a fixed point source anywhere in the troposphere is capable of being discussed by means of 2.02 with, under adiabatic conditions, $n = \frac{1}{4}$ and probably $C_x = C_y = C_z = 0.30 \text{ cm}^{\frac{1}{4}}$

## 3. Contrails

### (a) General

Aviation fuels are hydrocarbons which upon combustion with the oxygen of the air produce water and heat which are injected into and mix with the wake environment. As at very low temperatures the amount of water vapour which the air can hold in the form of invisible vapour is limited, at such temperatures a proportion of the water which is added appears in the form of visible droplets which then spread downwind and are scattered by turbulent diffusion.

In order to discuss the dispersion of these droplets, it will be assumed (1) that the evaporation of the droplets as they spread downwind can be neglected, and (2) that the turbulent diffusion from a point source moving at u cm. $\text{sec}^{-1}$ is the same as that from a fixed point source into a turbulent medium in which the mean wind speed is u cm. $\text{sec}^{-1}$. The limitations of these assumptions are discussed later in sections 4 and 5.

If therefore we take axes moving with the aircraft, with the origin at the exhaust with x measured downwind, y across wind and z in the vertical, then if $\varphi$ grams of fuel are used each second while cruising at u cm. $\text{sec}^{-1}$ and $\lambda$ grams of water are condensed as a result of the combustion of each gram of fuel, the amount of liquid water at x y z created by the passage of the aircraft is from 2.02

$$\Delta f = \frac{\varphi \lambda}{4\left[\Gamma\left(1+\frac{1}{1+m}\right)\right]^2 u\left(C^2 x\right)^{2/1+m}} \exp - \frac{|y|^{1+m} + |z|^{1+m}}{C^2 x} \qquad 3.00$$

where $C_x = C_y = C$, and $m = \frac{n}{2-n}$.

It is not altogether clear what criterion should be used for defining the visible edge of a contrail, but Roberts (2) in discussing the corresponding problem of a smoke trail assumed that a certain minimum number of particles in a tube of unit cross-section through the observer and the smoke cloud was required in order to obtain obscuration of the background.

/If

If we follow Roberts' opacity theory, then if N is this number and r and $\rho_d$ are the radius and density respectively of the droplets, assumed spherical,

$$\frac{4}{3}\pi r^3 \rho_d N = \int_{-\infty}^{\infty} \Delta f \, dy \qquad 3{\cdot}01$$

is the equation for the visible outline of the cloud.

Thus if $|z|$ is the visible radius of the trail at a distance x downwind,

$$\frac{4}{3}\pi r^3 \rho_d N = \frac{Q\lambda}{4\left[\Gamma\left(1+\frac{1}{1+m}\right)\right]^2 u (C^2 x)^{2/1+m}} \exp - \frac{|z|^{1+m}}{C^2 x} \int_{-\infty}^{\infty} \exp - \frac{|y|^{1+m}}{C^2 x} \, dy \qquad 3{\cdot}02$$

$$= \frac{Q\lambda}{2\Gamma\left(1+\frac{1}{1+m}\right) u (C^2 x)^{1/1+m}} \exp - \frac{|z|^{1+m}}{C^2 x} \qquad 3{\cdot}03$$

$$\therefore \frac{|z|^{1+m}}{C^2 x} = -\frac{1}{1+m} \log_e x - \log_e\left\{\frac{8}{3} u \Gamma\left(1+\frac{1}{1+m}\right) C^{\frac{2}{1+m}} \pi r^3 \rho_d N\right\} + \log_e (Q\lambda) \qquad 3{\cdot}04$$

If therefore we plot $|z|^{1+m}/x$ against $\log_e x$ with $m = \frac{1}{7}$ (which corresponds to $n = \frac{1}{4}$) we should obtain a straight line if the theory is at all in accord with the facts, and since the slope of the line is $C^2/(1+m)$, C can be determined without any assumptions as to the magnitudes of N, r or $\lambda$.

Alternatively if m is an index of the stability of the atmosphere at the level at which the aircraft is flying, then if $|z|^{1+m}/x$ is plotted against $\log_e x$ for different values of m, the points will only lie on a straight line for one particular value of m, so that photographs of contrails made under different conditions of stability appear to afford a possible method of determining m. 3.04 may also be written,

$$\frac{|z|^{1+m}}{x} = \frac{C^2}{(1+m)} \log_e \frac{L}{x} \qquad 3{\cdot}05$$

where L is the value of x for which $|z|$ vanishes, i.e. L is the length of the visible trail along the axis. It follows therefore that the trail length can be determined from measurements of the trail width at any point downwind if C and m are known and vice versa.

Differentiating 3.05 with respect to x and discarding the modulus we have

$$(1+m) z^m \frac{dz}{dx} = \frac{C^2}{1+m}\left[\log_e \frac{L}{x} - 1\right], \qquad 3{\cdot}06$$

so that if we consider a trail of given length L, the trail width is a maximum when

$$x = \frac{L}{2{\cdot}3} \qquad 3{\cdot}07$$

i.e. the trail attains its maximum width at about 3/4 of its length independently of C and m, cf. Fig. I which shows a good non-persistent contrail made by a Mosquito of the Meteorological Research Flight and photographed by a Mosquito of the Central Photographic Establishment, Benson.

The corresponding expression for the visible outline of a contrail from 2.01 is

$$\frac{z^2}{C^2 x^{2/(1+m)}} = \frac{1}{1+m} \log_e \frac{L}{x} \qquad 3{\cdot}08$$

so that observations of the radius of the contrail at specified distances downwind enable a comparison to be made between the two solutions 2.01 and 2.02.

## (b) The relation between trail length and drop size

If N is the number of droplets in a tube of unit cross-section through the observer perpendicular to and through the axis of the contrail required to obtain obscuration of the background, then from 3.03

$$\frac{4}{3}\pi r^3 \rho_d N = \frac{\phi\lambda}{2uT'\left(1+\frac{1}{1+m}\right)(C^2L)^{1/(1+m)}} \qquad 3{\cdot}09$$

and in obtaining this relationship no assumptions were made as to the magnitudes of N or r.

An approximate relationship between N and r may be obtained by the following method for which I am indebted to Dr. K.H. Stewart. If the angle subtended by the sun's radius at the earth is $\Theta$ radians and the apparent brightness of the sun is $B_1$ candles cm$^{-2}$, then the illumination produced by the sun on the contrail is $\pi B_1 \Theta^2$ lumens cm$^{-2}$. The brightness of the edge of the trail will thus be $(N A B_1 \Theta^2)/4$ candles cm$^{-2}$ where A is the effective scattering cross-section of the drop which, except when the angle of scattering is small, may be assumed to be equal to the geometrical cross-section $\pi r^2$.

If now the apparent brightness of the sky is $B_2$ candles cm$^{-2}$, the trail will be visible if the contrast between it and its background exceeds 0.02 (which is the value of the threshold of brightness used by Koschmieder (11)), so that N is given by

$$N r^2 = \frac{0{\cdot}08 B_2}{\pi B_1 \Theta^2} \qquad 3{\cdot}11$$

Using the values of the brightness of a clear blue sky $B_2 = 1{\cdot}2 \times 10^{-1}$ candles cm$^{-2}$, the brightness of the zenith sun $B_1$ = 1.65 x 10[illegible] candles cm$^{-2}$, and

$\theta = 4.665 \times 10^{-3}$ radians, (these values of $B_1$, $B_2$ and $\theta$ are taken from Tables of Physical and Chemical Constants by Kaye and Laby, the value of $B_2$ at 30,000 feet being reduced from that at the ground in the ratio of the pressures)

$$Nr^2 = 8.5 \times 10^{-4} \text{ cm}^2 \qquad 3.12$$

so that from 3.09

$$\frac{4}{3}\pi r \rho_d = \frac{\phi\lambda \times 1.05 \times 10^3}{2u\Gamma\left(1 + \frac{1}{1+m}\right)(C^2L)^{1/(1+m)}} \qquad 3.13$$

and hence from a knowledge of the trail length and $\lambda$ we may obtain the order of magnitude of r.

It follows from 3.09 that if $L_o$ is the length of trail for given values of $\varphi_o$, $u_o$ and $\lambda_o$ then the value of L for any other values of these variables is given by

$$L = L_o \left\{ \frac{\varphi \lambda u_o}{\varphi_o \lambda_o u} \right\}^{1+m} \qquad 3\cdot14$$

and since for straight and level flight at normal cruising speed the quantity of fuel used per second is directly proportional to the amount of air by weight which passes through the engine per second, it follows that the lengths of the trails at different levels should be related by

$$L = L_o \left\{ \frac{\lambda \rho}{\lambda_o \rho_o} \right\}^{1+m} \qquad 3\cdot15$$

where $\rho_o$ is the density of the air at the level where $L = L_o$ and $\lambda = \lambda_o$.

## (c) The value of $\lambda$.

It can be calculated from the chemical reactions involved that for each gram of fuel burned about 1.4 grams of water vapour and about 10,000 calories of heat are produced (1) so that if $\lambda$ grams appear in the form of visible water for each gram of fuel consumed then clearly $\lambda < 1.4$g.

An approximate expression for $\lambda$ which is convenient for numerical computation can be readily derived by a slight extension of an argument used in (1). If A is the cross-section of the trail into which the exhaust gases pass at a short distance downwind from the exhaust and it is assumed that all the fuel energy goes as heat into this trail [x], then the increase of temperature at this distance downwind is

$$\Delta T = \frac{10{,}000 \times \varphi}{A \rho c_p u} \qquad 3\cdot16$$

and similarly the increase in moisture content of $\Delta\mu$ grams per kilogram is

$$\Delta\mu = \frac{1\cdot4 \times 10^3 \times \varphi}{A \rho u} \qquad 3\cdot17$$

so that the ratio of the moisture increase to the temperature increase is independent of the cross-section of trail affected and is equal to 0.0336 gm kg$^{-1}$ per degree centigrade. This result which was also obtained by Appleman (12) shews that the passage of an aircraft through an initially saturated atmosphere would cause the atmosphere to become supersaturated if $0.0336 > d\mu_s/dT$, where $\mu_s$ is the saturated mixing ratio with respect to water corresponding to a temperature T, resulting in the production of water droplets.

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[x] The percentage of heat which is available for heating the air is usually assumed to be about 80% for piston driven aircraft and about 100% for jet aircraft, so that the values of $\lambda$ derived in this section may be too low. Provided observations from one type of aircraft only are used to compare theory and observation, errors from this assumption will not be important.

Now since when contrails are present the trail into which the exhaust gases are discharged is always saturated, it follows that the number of grams of water which are condensed for every gram of fuel burned is approximately

$$\lambda = \frac{A\rho U}{q \times 10^3}\left\{0{\cdot}0336 - \frac{d\mu_s}{dT}\right\}\Delta T \qquad 3{\cdot}18$$

or from 3.16

$$\lambda = \frac{10}{c_p}\left\{0.0336 - \frac{d\mu_s}{dT}\right\} \qquad 3{\cdot}19$$

so that from a knowledge of the variation with temperature of the saturation mixing ratio, $\lambda$ may be readily calculated from 3.19.

Values of $\mu_s$, the saturation mixing ratio over water, down to temperatures of -50°C, derived from the Goff-Gratch formula which is based upon the integration of the Clausius-Clapeyron equation and on modern experimental data, are given in the Smithsonian Meteorological Tables, but since as pointed out in these Tables, the necessary thermodynamical data for an exact integration of the Clausius-Clapeyron equation do not exist for temperatures below 0°C, the values of $\mu_s$ for these negative temperatures are open to doubt. A further difficulty results from the fact that contrails frequently form in the free air at temperatures down to -60°C and no values of $\mu_s$ are given for these temperatures. Values of $\mu_s$ for temperatures below -50°C have accordingly been derived by using the values of the saturation mixing ratio over ice for these temperatures, given in the Smithsonian Meteorological Tables, multiplied by the ratios, given in (1), of the saturation values of water vapour with respect to a water and ice surface respectively.

The values of $\lambda$ which have been derived from these values of $\mu_s$ and 3.19, and which may accordingly be subject to large errors, are shewn in Fig. II.

In order to simplify the problem it has up to the present been assumed that the contrail consists of water droplets which have formed immediately behind the aircraft in an environment which is saturated with respect to water, and that these droplets spread downwind without change of state. At the levels at which contrails occur the supercooled water droplets should in general freeze very quickly so that the resulting ice crystals would at first grow at the expense of the water environment. As the crystals spread downwind and began to mix with the air of the free environment, which in general is not saturated with respect to ice, they would commence to evaporate. This effect is briefly discussed in the following section.

## Evaporation of ice crystals in an unsaturated atmosphere

4. If we assume that the ice crystals are at rest relative to the air then as shewn by Jeffreys (13) the rate of evaporation of an ice crystal of mass $M$ in an unsaturated atmosphere is given by

$$\frac{dM}{dt} = 4\pi C_1 D(\rho_a - \rho_s) \qquad 4{\cdot}00$$

where $\rho_s$ is the vapour density at the surface of the crystal, $\rho_a$ is the vapour density at a point remote from the crystal, D is the coefficient of diffusion of water vapour in air and $C_1$ is the electrostatic capacity of the crystal, and this equation has been used to calculate the evaporation times of ice crystals whose shapes approximate to those of conductors of known capacity.

For a sphere of radius r, $C_1 = r$ and $M = \frac{4}{3}\pi r^3 \rho_i$, where $\rho_i$ is the crystal density, so that 4.00 may be written

$$r\frac{dr}{dt} = \frac{D}{\rho_i}\left(\rho_a - \rho_s\right) \qquad 4{\cdot}01$$

If therefore r is the radius of the crystal at a time t and $r_o$ is the radius of the crystal initially, then

$$r^2 = r_0^2 - \frac{2tD}{\rho_i}\left(\rho_s - \rho_a\right) \qquad 4{\cdot}02$$

It follows therefore that the time taken for a spherical crystal of radius r to evaporate completely is

$$t_1 = \frac{\rho_i r_0^2}{2D\left(\rho_s - \rho_a\right)} \qquad 4{\cdot}03$$

i.e. the drier the environment and the smaller the ice crystal, the more rapidly does it evaporate.

The capacity of an ellipsoidal conductor differs only slightly from that of a spherical conductor of the same volume, and if the axes of the ellipsoid are 2r $(1+\alpha)$, 2r $(1+\beta)$, 2r $(1+\gamma)$ where $(1+\alpha)(1+\beta)(1+\gamma) = 1$, the capacity is approximately

$$C_1 = r\left[1 + \frac{2}{15}\left(\alpha^2 + \beta^2 + \gamma^2\right)\right] \qquad 4{\cdot}04$$

Initially therefore an ice crystal whose shape is ellipsoidal would commence to evaporate at a slightly faster rate than a spherical ice crystal of the same volume. In the case of an ellipsoidal crystal however the evaporation would be greatest at the ends of the major axis and least at the ends of the minor axes and would thus tend to promote spherical symmetry and thereby reduce the differential rate of evaporation, so that the evaporation times of an ellipsoidal crystal should be of the same order as those of a spherical crystal.

For simplicity we shall assume that the ice crystals may be treated as spheres, so that from 4.02 and 4.03

$$\frac{r^2}{r_0^2} = 1 - \frac{t}{t_1}, \qquad t_1 > t \qquad 4{\cdot}05$$

This equation is only valid at some distance from the exhaust when the ice crystals have become widely separated. Now from 3.03 the number of ice crystals lying in a tube of unit cross-section through the observer and perpendicular to and through the axis of the contrail varies as $x^{-1/(1+m)}$, where x is the distance downwind from the aircraft. Hence since the obscuring power varies as the number of crystals and the square of their radii, it follows that if L is the visible length of the trail with particles of radius $r_o$ and $L_1$ is the length of trail with particles of radius r, then

$$\frac{r^2}{r_0^2} = \frac{L_1^{\frac{1}{1+m}}}{L^{\frac{1}{1+m}}} \qquad 4{\cdot}06$$

and hence combining 4.05 and 4.06

$$L_1 = L\left\{1 - \frac{t}{t_1}\right\}^{1+m} \qquad 4{\cdot}07$$

or since the time an ice crystal has been in existence at a distance $L_1$ downwind from an aircraft moving at u cm. $sec^{-1}$ is $L_1/u$, then

$$L_1 = L\left\{1 - \frac{L_1}{ut_1}\right\}^{1+m} \qquad 4{\cdot}08$$

## 5. Comparison with experiments

### (a) Length and occurrence of contrail

In a conjoint investigation by the Meteorological Research Flight, Farnborough, in cooperation with the Central Photographic Establishment, Benson, two Mosquito aircraft, one from each unit flew in close proximity, the observer in the M.R.F. aircraft taking observations of the temperature and frost point and the observer in the C.P.E. aircraft taking photographs of the contrails made by the M.R.F. aircraft. The observer in the M.R.F. aircraft also made estimates of the length, intensity and persistence of the contrails made by the C.P.E. aircraft.

Table I gives details of the observations made by the M.R.F. aircraft on March 11th, 1952, whilst Table II gives details of the radii of the M.R.F. contrail at various distances downwind as calculated from photographs made by the C.P.E. aircraft.

The observations were made in the upper troposphere in returning maritime polar air over the Isle of Wight.

During the period of observation, the weather was fine with only small amounts of cumulus cloud of limited vertical extent and with base at 2,500 feet. The winds which were less than 15 kt. at all levels were from WNW between 30,000 and 35,000 feet, and the tropopause which was at 36,000 feet coincided with the maximum heights attained by the aircraft.

Table I

Observations of temperature and frost point on 11th March, 1952

| ICAN heights in ft. | Pressure mb. | Temperature °F | Frost Point °F |
|---|---|---|---|
| 27000 | 340 | -43° | -55° |
| 28000 | 329 | -48 | -55 |
| 29000 | 315 | -52 | -66 |
| 30000 | 301 | -56 | -68 |
| 31000 | 287 | -62 | -71 |
| 32000 | 274 | -65 | -74 |
| 33000 | 262 | -69 | -76 |
| 34000 | 250 | -73 | -80 |
| 35000 | 238 | -77 | -81 |
| 36000 | 227 | -81 | -88 |
| 38500* | 200 | -71 | - |
| 44500* | 150 | -64 | - |
| 53200* | 100 | -61 | - |

* Values taken from the Larkhill Radiosonde Ascent at 1500 G.M.T. on March 11th.

/Table II

## Table II

### Radius of the trail at specified distances downwind from the aircraft

| Radius of trail in m. | Distance downwind from aircraft in m. |
|---|---|
| 1.05 | 12 |
| 1.50 | 24 |
| 1.80 | 36 |
| 1.95 | 48 |
| 2.10 | 60 |

Curve A, Fig. III shews the results of plotting $z^{8/7}/x$ against $\log_{10} x$ and it can be seen that as required by 3.05 the points lie on a straight line from whose slope a value of $C_z = 0.26$ is obtained in very good agreement with the values of $C_y$ and $C_x$ derived from Jones' smoke puff observations. Curve B, Fig. III shews the results of plotting $z^2/x^{7/4}$ against $\log_{10} x$ in order to test out 3.08 and it can be seen that in this case the points do not conform to a straight line.

The intercept of curve A with the log x axis gives the visible length of the trail downwind as 215 m. It was unfortunately not possible on this occasion to check this from calculations of the length of the M.R.F. trail from the C.P.E. photographs, as in order to obtain accurate measurements of the trail radius the aircraft flew too close for complete trail lengths in excess of 70 m. to be photographed.

The length of the C.P.E. contrail however as estimated by the observer in the M.R.F. aircraft cf. Table IV was 150 m. which is in reasonably good agreement with, and as expected from 4.08 is lower than, the theoretical value. The effect of the humidity of the environment on the trail length is discussed in 5(c).

Curve a, Fig. II shews the temperature observations for March 11th, 1952, plotted on a height temperature diagram supplemented by temperatures above 36,000 feet taken from the Larkhill ascent on the same day. It can be seen from this curve that if our computations of $\lambda$ are correct, no liquid water can be condensed below 26,000 feet or above 50,000 feet, so that on this occasion contrails could have been predicted as liable to form at all levels between 27,000 feet and 49,000 feet. It can also be seen from this curve that the value of $\lambda$ at 30,000 feet is 0.73 g. At 30,000 feet, the aircraft flew at reduced speeds varying from a true airspeed of 235 kt. to 290 kt. while the fuel consumption varied from 80 gallons per hour to 110 gallons per hour, so that $Q/u$ varied from $6.71 \times 10^{-3}$ g. $cm^{-1}$ at the lowest speed to $7.46 \times 10^{-3}$ g. $cm^{-1}$ at the highest speed. The value of $Q/u$ at 30,000 feet may therefore be taken as $7.1 \times 10^{-3}$ g. $cm^{-1}$ with a tolerance of about ± 5%.

It follows from 3.00 that the value of $\delta f$ at the visible edge of the contrail at a distance of 215 m. downwind is

$$\delta f = \frac{7 \cdot 1 \times 10^{-3} \times 7 \cdot 3 \times 10^{-1}}{4\left[\Gamma(1\tfrac{7}{8})\right]^2 \left[(0 \cdot 26)^2 \times 21500\right]^{7/4}}$$

$$= 4 \cdot 14 \times 10^{-3} \text{ g m}^{-3} \qquad 5.00$$

No direct measurements of the amount of solid water in contrails are available to check this, but Goldie (14) and Parker (15) have estimated that the least density of water in ice crystal form required to produce a visible contrail should be of the order of $4 \times 10^{-3}$ g. $m^{-3}$ in good agreement with this theoretical result.

/Table III

Table III shews the lengths of contrails as computed from 3.15 together with details of the trails as estimated and photographed. Since the values of $Q/\upsilon$ may be in error by about 5% it follows from 3.14 that the trail lengths may be in error by about the same amount due to this cause. The neglect of evaporation of the ice crystals as they spread downwind in an unsaturated atmosphere leads to a slight overestimation of the trail lengths, but as will be shewn later in 5(c) this effect is in general small. The errors in $\lambda$ are unknown but, from a posteriori considerations, appear to be small.

Table III - Computed and observed contrails 11th March, 1952

| Height in feet | $\lambda$ g. | $\frac{Q}{\upsilon}$ x$10^{-3}$ g. sec$^{-1}$ | L computed in m. | Details of MRF trail from photographs | Details of CPE trail as estimated |
|---|---|---|---|---|---|
| 27000 | 0.16 | 7.92 | 43 | weak trail > 67 m. | slight trail 48 m. |
| 28000 | 0.38 | 7.64 | 111 | weak trail > 67 m. | slight trail 48 m. |
| 29000 | 0.55 | 7.37 | 162 | moderate trail > 67 m. | moderate trail 111 m. |
| 30000 | 0.73 | 7.10 | 215 | good trail > 67 m. | moderate trail 150 m. |
| 31000 | 0.87 | 6.83 | 251 | good trail > 67 m. | long trail |
| 32000 | 0.93 | 6.59 | 260 | good trail > 67 m. | long trail |
| 33000 | 1.01 | 6.35 | 274 | good trail > 67 m. | long trail |
| 34000 | 1.07 | 6.12 | 280 | good trail > 67 m. | long trail |
| 35000 | 1.12 | 5.90 | 284 | good trail > 67 m. | long trail |
| 36000 | 1.17 | 5.67 | 285 | good trail > 67 m. | long trail |
| 38500 | 0.93 | 5.14 | 196 | No observations | |
| 44500 | 0.43 | 4.07 | 90 | No observations | |
| 53200 | 0 | - | 0 | No observations | |

Two other series of double flights made by the M.R.F. and C.P.E. Mosquitos are also available for discussion and these have been plotted as curves b and c in Fig. II.

It can be seen from curve b, which shews the temperatures for March 4th, 1952, that no liquid water can condense below 24,000 ft. or above 34,000 ft. so that on this occasion contrails could have been predicted as liable to form between these limits. These in point of fact materialised and, together with the computed values, are detailed in Table IV.

Table IV - Computed and observed contrails on 4th March, 1952

| Height in feet | $\lambda$ g. | $\frac{Q}{\upsilon}$ x$10^{-3}$ g. cm$^{-1}$ | L computed | Details of MRF trail from photographs | Details of CPE trail as estimated |
|---|---|---|---|---|---|
| 25000 | 0.30 | 8.49 | 97 | No trails visible | 25 very faint |
| 26000 | 0.50 | 8.20 | 166 | 60 moderate | 25 faint |
| 27000 | 0.60 | 7.92 | 196 | 90 moderate | 39 moderate |
| 28000 | 0.54 | 7.64 | 167 | 150 moderate | 48 moderate |
| 29000 | 0.70 | 7.37 | 214 | 195 dense | 87 moderate |
| 30000 | 0.60 | 7.10 | 172 | 210 dense | 63 moderate |
| 31000 | 0.57 | 6.83 | 157 | 150 moderate | 63 moderate |
| 32000 | 0.25 | 6.59 | 58 | intermittent trail | 39 moderate |
| 33000 | 0.20 | 6.35 | 42 | 60 weak | 30 moderate |
| 34000 | 0 | 6.12 | 0 | No trail visible | No trail visible |

The descriptive comments of the contrails made by the C.P.E. aircraft and noted by the M.R.F. observer were very consistent with deductions made from photographs of the contrails made by the M.R.F. aircraft, but for the flights on March 4th, 1952, cf. Table IV the estimated trail lengths were consistently lower than the measured trail lengths.

This may be due to a systematic tendency for visual estimates to be less than measurements from photographs, due to differences in threshold contrasts, or to the difference in view with respect to the sun with consequently very different illumination, or to the trails from the two aircraft being of very different lengths, as on this occasion the photographed M.R.F. aircraft was overtaking the photographing C.P.E. aircraft and was therefore under more power and generating more trail cf. 3.14, or to a combination of all three causes.

It can be seen from curve c which depicts the temperatures for October 24th, 1952, that no liquid water can condense out at any level and hence 'Nil contrails' could have been forecast and none in fact occurred.

Inspection of Fig. II in conjunction with these three ascents suggests that the isopleth $\lambda = 0$ which gives temperature limits below which contrails form and above which contrails are impossible has been correctly derived, and Fig. IV which shews the first appearances of all contrails together with details of the relative humidity with respect to ice of the free environment as recorded by the M.R.F. between the 30th November, 1951, and March 13th, 1953, support this. It can also be seen from Fig. IV that the first appearance of the contrail bears no relationship to the humidity of the environment, a fact which was first noted by Bannon (16).

## (b) Crystal size and persistence of contrail

Using the values of $\frac{q}{v} = 7.1 \times 10^{-3}$ g. cm$^{-1}$, $\lambda = 7.3 \times 10^{-1}$ g. and L = 215 m. which were found at 30,000 feet on March 11th, 1952, the radius of the ice crystals, assumed spherical, is from 3.13

$$r = \left[\frac{7.1 \times 10^{-3} \times 7.3 \times 10^{-1} \times 1.05 \times 10^{3}}{\frac{4}{3} \times \frac{22}{7} \times 9.2 \times 10^{-1} \times 2 \times T(1\tfrac{7}{8}) \times \left[(0.26)^2 \times 21500\right]}\right]^{7/8}$$

$$= 14\mu \qquad\qquad 5.01$$

That this is of the right order of magnitude is shewn by laboratory experiments by Mason (17) on the rate of growth of ice crystals in a supersaturated atmosphere which indicated that at temperatures of about -40°C the ice crystals occurred mainly in the form of hexagonal prisms whose equivalent radii were about 13$\mu$ to 20$\mu$, and the observations of ice crystals in the single penetration into a long and persistent contrail made by an aircraft of the M.R.F. in general support this. ¤

Using the data of March 11th, 1952, which is given in Table II, the times taken for a spherical crystal of radius 15$\mu$ to evaporate completely have been computed from 4.03 and these times together with details of the persistency of the trails noted by the M.R.F. observer are given in Table V.

Table V - Evaporation times and persistency 11th March, 1952

| Height | $\rho_s - \rho_a$ | $t_1$ sec. | Persistency = P<br>Non Persistency = N.P. |
|---|---|---|---|
| 27000 | .052 | 110 | N.P. |
| 28000 | .026 | 220 | N.P. |
| 29000 | .034 | 170 | N.P. |
| 30000 | .023 | 240 | N.P. |
| 31000 | .013 | 420 | N.P. |
| 32000 | .011 | 490 | P. |
| 33000 | .007 | 790 | P. |
| 34000 | .006 | 920 | P. |
| 35000 | .003 | 1850 | P. |
| 36000 | .004 | 1400 | P. |
| 38500 | (.040) | (14) | (N.P.) |

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¤ Communicated privately by Mr. R.J. Murgatroyd of M.R.F.

In the stratosphere the frost point usually decreases rapidly with height while the temperature either remains steady or increases with height so that in the stratosphere $\rho_s - \rho_a$ increases with height cf. Bannon and others (16). The estimates of $\rho_s - \rho_a$ and of the persistency at 38,500 are typical of conditions in the stratosphere with a well marked Type I tropopause at about 36,000 feet.

Table VI gives the temperatures and frost points for the ascents on March 4th together with the theoretical evaporation times for a spherical crystal of radius 15 $\mu$ computed from 4.03 and details of the persistency as noted by the M.R.F.

Table VI - Evaporation times and persistency March 4th, 1952

| Height | Temperature °F | Frost Point °F | $\rho_s - \rho_a$ | $t_1$ sec. | Persistency |
|---|---|---|---|---|---|
| 25000 | -44 | -56.5 | .053 | 110 | N.P. |
| 26000 | -48 | -68 | .054 | 105 | N.P. |
| 27000 | -51 | -71 | .045 | 125 | N.P. |
| 28000 | -51 | -78 | .051 | 115 | N.P. |
| 29000 | -55 | -83 | .041 | 140 | N.P. |
| 30000 | -54 | -87 | .044 | 130 | N.P. |
| 31000 | -54 | -81.5 | .042 | 135 | N.P. |
| 32000 | -49 | -93 | .061 | 95 | N.P. |
| 33000 | -49 | -94 | .065 | 85 | N.P. |

Inspection of Tables V and VI suggests that there is a general relationship between these theoretical evaporation times and the persistency, and that for theoretical evaporation times (with ice crystals initially of radius 15 $\mu$ ) of about 7 minutes or less, contrails are non persistent.

(c) Persistency and trail length

Since in the preceding analysis only qualitative use has been made of the theoretical evaporation times of a spherical ice crystal, it makes no difference whether the times in Tables V and VI are as given, or whether they should be multiplied by some constant factor to take account of the shape effect. In the present section however in which we are concerned with the actual evaporation times, it is necessary to consider whether the shapes in which ice crystals are normally found, i.e. flat hexagonal plates or hexagonal prisms, materially alter the order of magnitude of these times.

The capacities of hexagonal plates and prisms are difficult to calculate, but Mason (17) found that to a first approximation hexagonal plates could be treated as oblate spheroids and hexagonal prisms as prolate spheroids.

It can be seen from 4.04 that the capacities of an oblate spheroid and of a prolate spheroid are approximately the same and if, to get some idea of the magnitude of the shape effect, we take $\alpha = \frac{1}{2}$, $\beta = 0$ and $\gamma = -\frac{1}{3}$, which corresponds to the case of an oblate spheroid in which the major axis is approximately double the minor axis, $C = 1.06\ r$ approximately. Hence a crystal originally in the shape of an oblate spheroid will commence to evaporate slightly faster than a spherical crystal of the same volume, and the evaporation times of the former will be lower than those of the latter by about 3%. It would thus appear that the theoretical evaporation times of a spherical ice crystal may be taken as representing to a first approximation the actual evaporation times of ice crystals of equal volume but of different shapes.

Table VII which has been constructed from 4.08 shews how contrails of various lengths are reduced for specified evaporation times and it can be seen from this table that for contrails below 500 m. in length the evaporation from the ice crystals as they spread downwind has very little effect upon the trail length.

Table VII - Evaporation times and trail lengths

| t<br>L | ∞ | 400 | 40* |
|---|---|---|---|
| 500 | 500 | 495 | 449 |
| 250 | 250 | 249 | 236 |
| 200 | 200 | 199 | 192 |
| 100 | 100 | 100 | 99 |
| 50 | 50 | 50 | 50 |

For trail lengths in excess of some particular value which, in the case of Mosquitos flying under reduced power, is about 250 m. cf. Table III, the predicted lengths based upon an infinite evaporation time are less than those actually observed, or in other words the rate of diffusion of ice crystals at distances in excess of this length is far slower than that predicted by 2.03.

This is probably a consequence of the assumption that the trail from an aircraft moving through the air with a velocity u cm. $sec^{-1}$ is the same as that from a fixed point in the atmosphere in which the mean wind speed is u cm. $sec^{-1}$. It is clear that this assumption can only be approximately correct for a limited distance downwind, for in the immediate wake of the aircraft we have found that the diffusion coefficients are represented by expressions similar to those found from Jones' smoke puff observations at points remote from the aircraft with however u the velocity of the aircraft relative to the air in the former replacing U the velocity of the air in the latter. It suggests however that the turbulent wake behind a piston type aircraft extends for about 250 metres and that if the turbulence in the wake is sufficient to dissipate the contrail before this length is attained then only non persistent contrails are formed. It follows therefore that a necessary condition for a persistent contrail is $\frac{q\lambda}{U} > 5 \times 10^{-3}$.

## Conclusions

A comparison between contrail observations made conjointly by Mosquitos of the Meteorological Research Flight and the Central Photographic Establishment and deductions made from the solution of the continuous point source problem, together with certain simplifying assumptions, indicates

(1) that the lengths and widths of non persistent contrails can be predicted with considerable accuracy from a knowledge of the fuel consumption, speed of aircraft and temperature of the environment alone,

(2) that the formation of contrails is independent of the humidity of the air and depends only upon the temperature of the environment and to a slight extent upon the type of aircraft fuel,

(3) that contrails form anywhere within a zone given by the intersection of the environment curve and the isopleth $\lambda = 0$ and increase in length up to the level of the tropopause, if this is within the zone, and decrease very rapidly with height in the stratosphere

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* Unlikely to be encountered

and suggests

(4) that long and persistent trails should only occur at temperatures below -40° in (a) an atmosphere which is supersaturated with respect to ice and for which $\lambda > 0$, so that any droplets which are produced would freeze and grow at the expense of the environment or (b) at very low temperatures where $\frac{Q\lambda}{u} > 5 \times 10^{-3}$ together with a sufficiently high relative humidity with respect to ice for the evaporation time of a spherical ice crystal of radius 15$\mu$ to be greater than 7 minutes.

A preliminary examination of long and persistent contrails made by Mosquitos of the M.R.F. shews that nearly 90% fall within categories (a) and (b) but the remaining 10%, if the observing and instrumental techniques of humidity measurements by aircraft are correct, indicate that some other factor which has not been taken into account in the foregoing analysis is required.

Acknowledgment is due to Mr. R.J. Murgatroyd, Senior Meteorological Officer, Meteorological Research Flight, Farnborough and Mr. G.W. Hurst, Senior Meteorological Officer, Central Photographic Establishment, Benson for the observations and photographs of contrails on which this report is based.

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| (17) | B.J. Mason | 1953 | 'The growth of ice crystals in a supercooled water cloud'. Q.J.R.M.S. 79, p.104. |

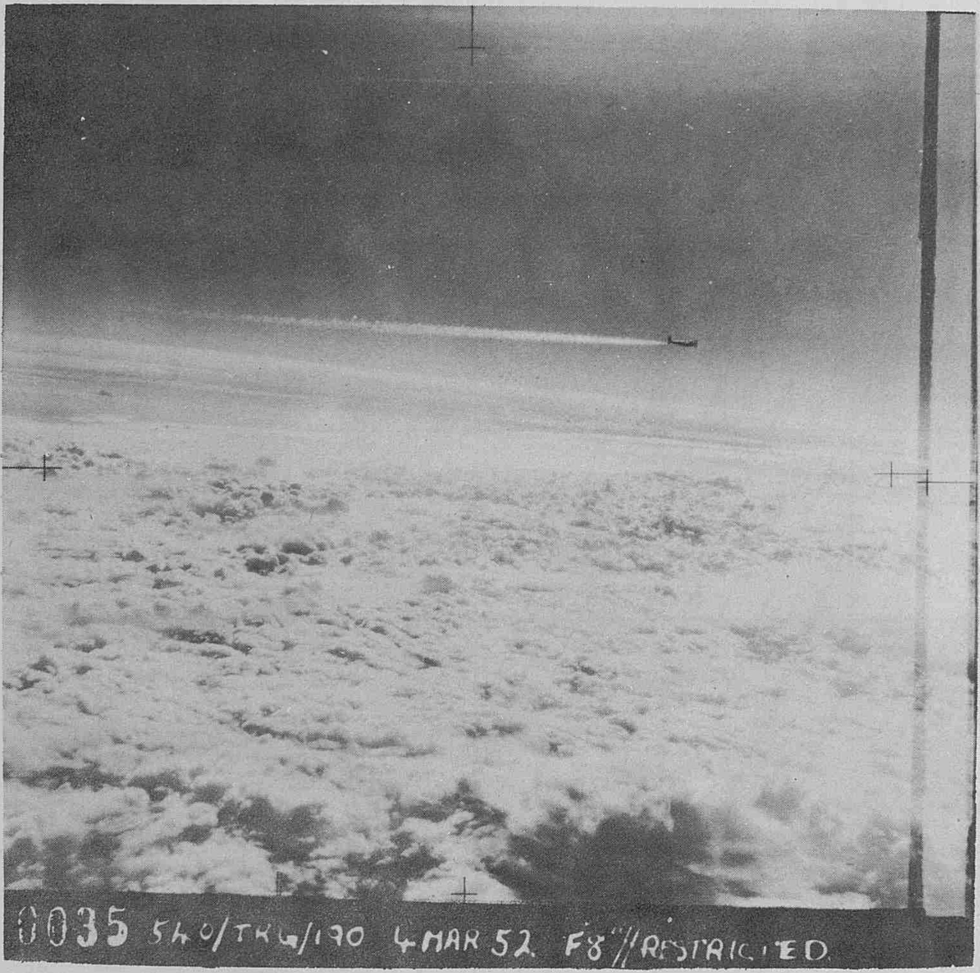
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Fig 1. Non-persistent contrail, Larkhill area.

M.R.P. 830.

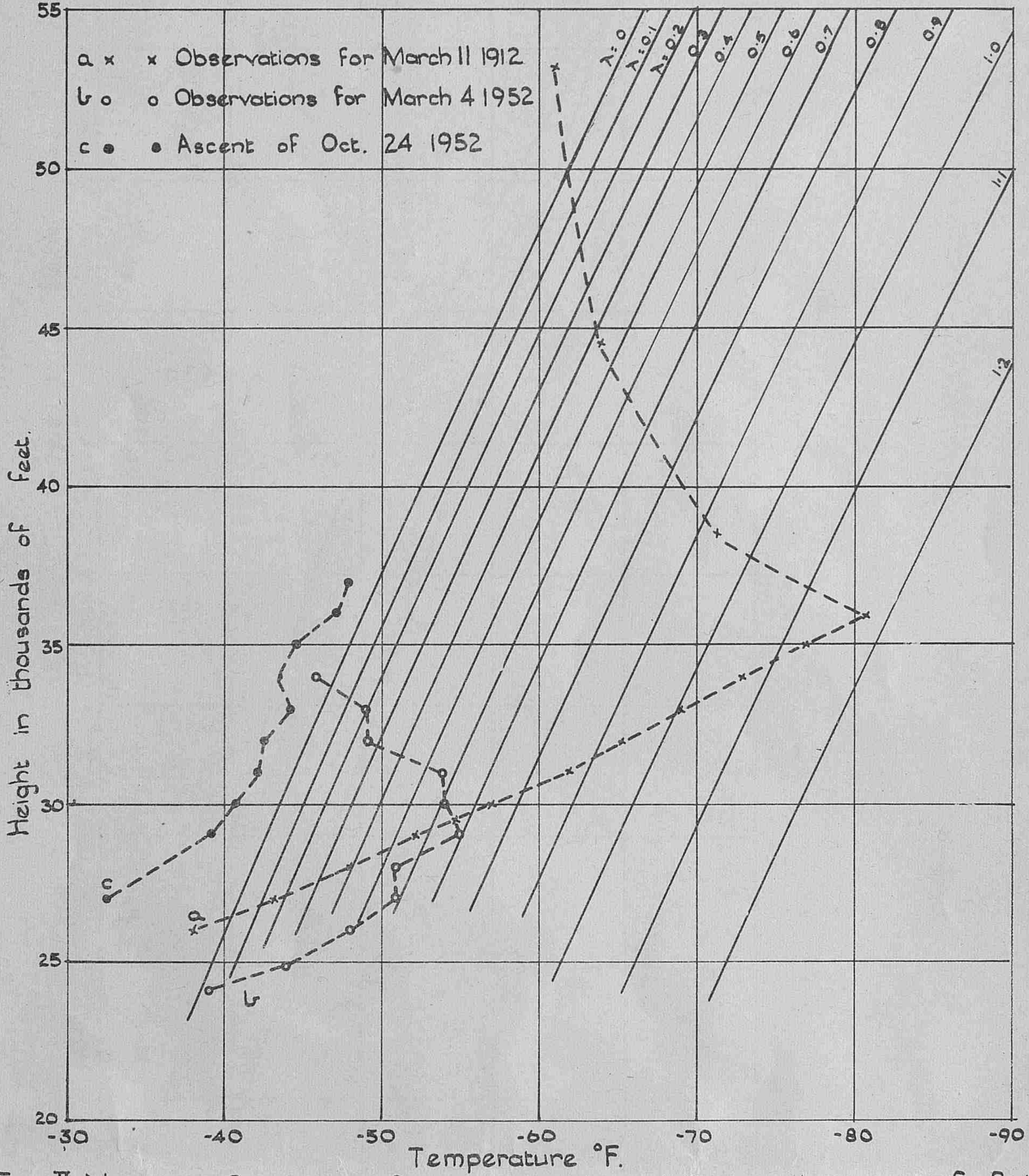

Fig. II. Number of grams of water condensed for each gram of fuel burned.

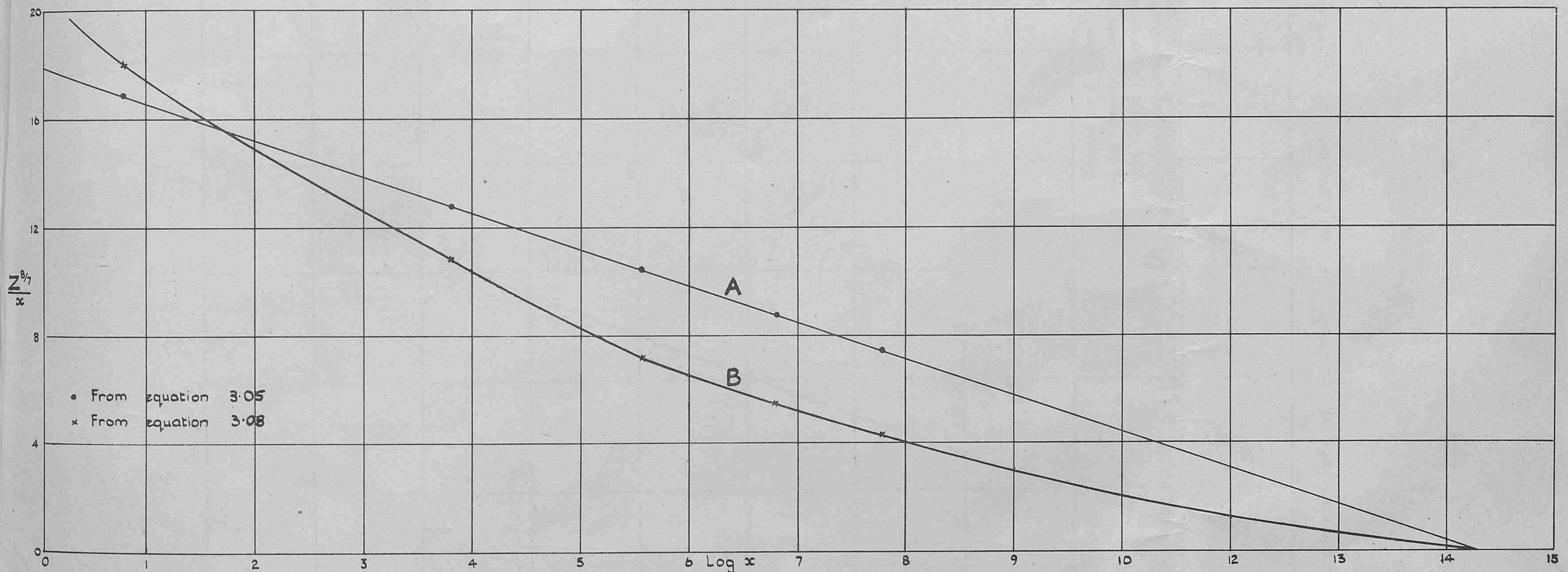

Fig. III. Growth of the visible outline of a contrail.

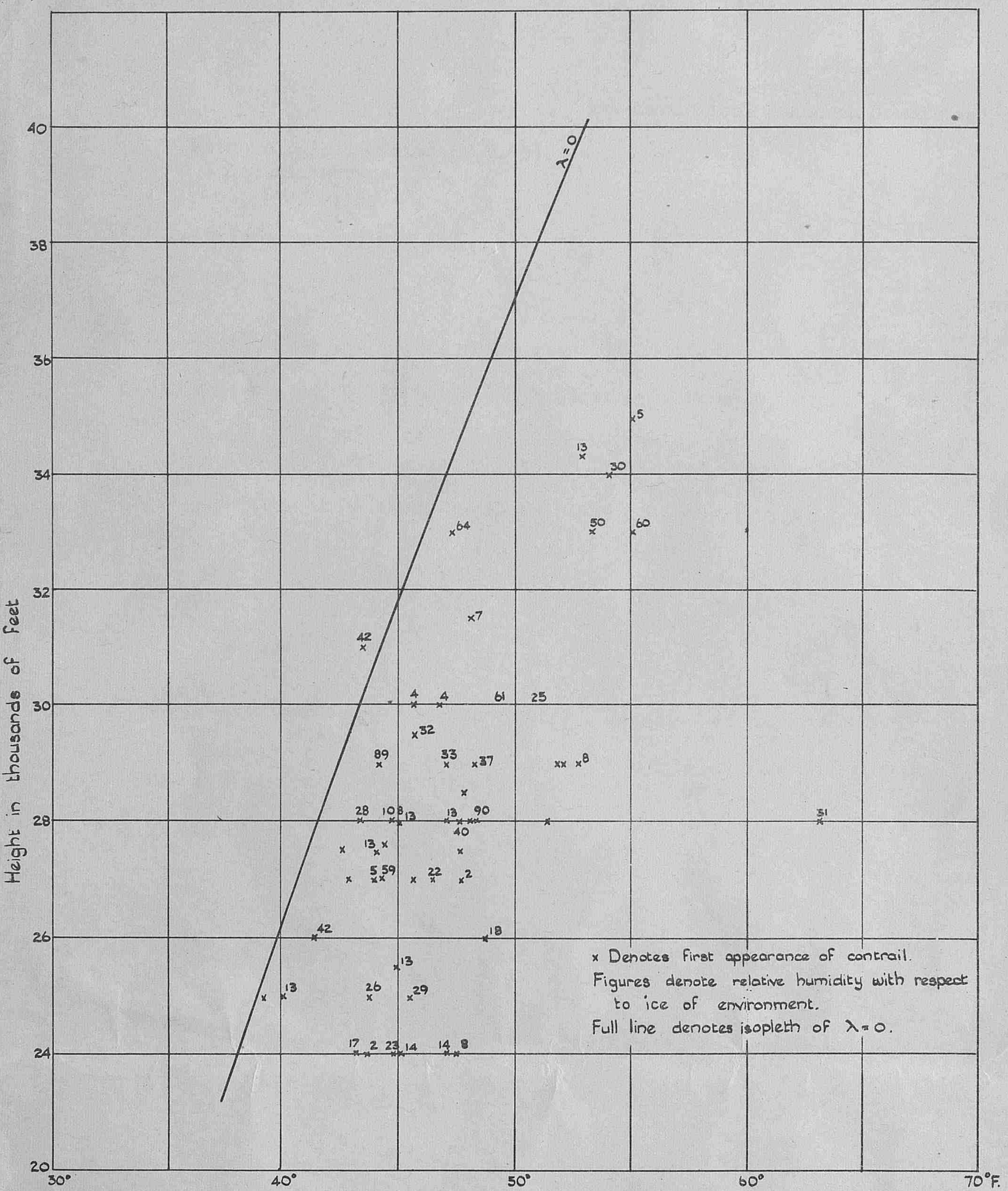

Fig. IV First appearance of a contrail as a function of height and temperature.

M.O.15/R/7

To all recipients of M.R.P. 830

Will you please attach the enclosed addendum to your copy of M.R.P. 830 and also note that this paper has now been declassified.

H W L Absalom
Secretary,
Meteorological Research Committee

Meteorological Office (M.O.15),
Air Ministry,
London, W.C.2.

22nd February, 1957.

G.5280/PG/2/57/170

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is assumed that the mixing takes place in the turbulent wake immediately behind the aircraft and that the droplets resulting from this mixing are ejected from the end of this wake into the free atmosphere where they diffuse normally.

Thus if we take axes moving with the aircraft, with the origin at the points where the droplets are ejected from the wake, then with x measured downwind, y across wind and z vertically upward the equation which states that the advection is balanced by lateral and vertical diffusion is as in 2.04

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left(K_y\frac{\partial \chi}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial \chi}{\partial z}\right) \qquad 6.01$$

where U is the velocity of the aircraft relative to the air.

Now in the case in which diffusion takes place immediately behind the aircraft the expressions used for the coefficients of eddy diffusion in the y and z directions respectively were

$$K_y = \frac{UC^2|y|^{1-m}}{(1+m)^2} \qquad 6.02$$

and $$K_z = \frac{UC^2|z|^{1-m}}{(1+m)^2} \qquad 6.03$$

and the solution of equation of eddy diffusion

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left\{\frac{UC^2|y|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial y}\right\} + \frac{\partial}{\partial z}\left\{\frac{UC^2|z|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial z}\right\} \qquad 6.04$$

with the appropriate boundary conditions, was used to discuss the problem.

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Weather Service

## Addendum to M.R.P. 830

## The prediction of contrails from jet aircraft

In para. 5(c) of M.R.P. 830 it was noted that for trail lengths in excess of about 250 metres the predicted lengths were less than those observed and it was suggested that this was a consequence of the assumption that the trail from an aircraft moving through the air with a velocity of U cm $\sec^{-1}$ is the same as that from a fixed point in the atmosphere in which the mean speed is U cm $\sec^{-1}$. It was pointed out that this assumption could only be true for a limited distance downwind from the aircraft for in the immediate turbulent wake it was found that the diffusion coefficients were the same as those in the free air apart from the fact that in the former U the velocity of the aircraft relative to the air replaced V the mean wind speed in the latter, and it was suggested that the turbulent wake extended for about 250 metres.

With a jet aircraft in which trail lengths are of the order of a mile or more and where a knowledge of the outline of the contrail in the vicinity of the exhaust is of little significance, the problem of predicting the occurrence, length, width and persistence of contrails can be dealt with very simply if it is assumed that the mixing takes place in the turbulent wake immediately behind the aircraft and that the droplets resulting from this mixing are ejected from the end of this wake into the free atmosphere where they diffuse normally.

Thus if we take axes moving with the aircraft, with the origin at the points where the droplets are ejected from the wake, then with x measured downwind, y across wind and z vertically upward the equation which states that the advection is balanced by lateral and vertical diffusion is as in 2.04

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left(K_y \frac{\partial \chi}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial \chi}{\partial z}\right) \qquad 6.01$$

where U is the velocity of the aircraft relative to the air.

Now in the case in which diffusion takes place immediately behind the aircraft the expressions used for the coefficients of eddy diffusion in the y and z directions respectively were

$$K_y = \frac{UC^2|y|^{1-m}}{(1+m)^2} \qquad 6.02$$

and $$K_z = \frac{UC^2|z|^{1-m}}{(1+m)^2} \qquad 6.03$$

and the solution of equation of eddy diffusion

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left\{\frac{UC^2|y|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial y}\right\} + \frac{\partial}{\partial z}\left\{\frac{UC^2|z|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial z}\right\} \qquad 6.04$$

with the appropriate boundary conditions, was used to discuss the problem.

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In the present case using coefficients of eddy diffusion appropriate to the free air, i.e., with V replacing U, the equation to be solved is

$$U\frac{\partial \chi}{\partial x} = \frac{\partial}{\partial y}\left\{\frac{VC^2|y|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial y}\right\} + \frac{\partial}{\partial z}\left\{\frac{VC^2|z|^{1-m}}{(1+m)^2}\frac{\partial \chi}{\partial z}\right\} \qquad 6.05$$

with the same boundary conditions as before.

If however we transform this equation by introducing a new coefficient $C_1$ where

$$UC_1^2 = VC^2 \qquad 6.06$$

then equations 6.05 and 6.04 are the same with $C_1$ replacing C and hence any solutions derived and used in M.R.P. 830 can be used in the present discussion providing $C^2$ wherever it occurs is now replaced by $\frac{VC^2}{U}$ . For example the visible outline of a contrail immediately behind a piston type aircraft is from 3.05,

$$\frac{|z|^{1+m}}{x} = \frac{C^2}{1+m}\log_e\frac{L}{x} \qquad 6.07$$

so that the visible outline of a contrail at some distance behind a jet aircraft is given by

$$\frac{|z|^{1+m}}{x} = \frac{VC^2}{U(1+m)}\log_e\frac{L}{x} \qquad 6.08$$

and again the trail width is a maximum at about 3/7 of its length.

Similarly 3.13 which gives the relationship between the trail length L, drop radius r, fuel consumption per cm of path $\frac{\Phi}{U}$ and amount of water which is condensed as the result of combustion of each gram of fuel $\lambda$ , shows that for the piston type aircraft,

$$L = \frac{\text{const.}}{C^2}\left(\frac{\phi\lambda}{Ur}\right)^{1+m} \qquad 6.09$$

so that in the case of jet aircraft the relationship becomes

$$L = \frac{\text{const.}\ U}{C^2 V}\left(\frac{\phi\lambda}{Ur}\right)^{1+m} \qquad 6.10$$

According to details given by Bannon (18) the fuel consumption per cm of path at 30,000 feet for a Meteor IV is approximately the same as that of the Mosquito, details of whose contrails are given in Table IV.

If therefore $\lambda$ is the same in both cases then with a speed of 500 kt a Meteor IV would have produced a visible trail of approximately 2,000 metres, if the wind speed had been 50 kt, as opposed to the visible trail of approximately 200 metres produced by the Mosquito.

Reference

(18) J.K. Bannon 1949 Meteorological Research Papers, Air Ministry, London, No. 489.

C.D. 8053

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 831 by Mr. M.J. Blackwell. A copy is available in the Library of the Meteorological Office."

M.R.P. 831
S.C.III/162
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# AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

## Five years continuous recording of daylight illumination at Kew Observatory

by

M.J. Blackwell (Meteorological Office)

## Summary

The intensity of illumination on a horizontal surface has been recorded continuously at Kew Observatory since January, 1947. In this paper, the various errors inherent in the measurements are considered; the methods of correcting the basic tabulations are demonstrated; and the corrected data are presented in the form of hourly means for 10 day periods.

Examples of the diurnal variation of illumination on cloudless and overcast days are given, and the frequency distribution of illumination at noon for selected months is indicated. The relation between illumination and solar (sun + sky) radiation is discussed.

## 1. Introduction

(a) Some remarks on photometry. The object of the work to be described is the continuous measurement of the luminous flux through a horizontal surface fully exposed to sun and sky, by instrumental photometry of a white diffusing surface. The illumination of a surface is measurable by visual comparison with a similar surface illuminated by a standard source. The ultimate standard of luminous intensity is a full (black body) radiator at the temperature of solidifying platinum (1), (2) but in practice it is replaced by means of standard lamps.

A standard eye has been specified by the International Commission on Illumination from the results of experiments on numerous observers (3). This specification relates the relative sensitivity of the eye, throughout the spectrum, to a given quantity of monochromatic radiation. Although the sensitivity function varies with intensity at lower levels of illumination, it is constant in the range with which this paper is concerned (photopic vision). This puts the comparison of illumination due to sources of different colours on a sound basis, and facilitates instrumental rather than visual photometry, it being merely necessary to ensure that the relative response of the instrument is the same as that of the I.C.I. eye throughout the spectrum.

The unit of illumination is the lux, the illumination produced by a source of unit candle power (1 candela) at a distance of 1 metre, where 1 [illegible] under a jet aircraft is given by

$$\frac{(z)^{1+m}}{x} = \frac{VC^2}{U(1+m)} \log_e \frac{L}{x} \qquad 6.08$$

and again the trail width is a maximum at about 1/3 of the length.

Similarly 3.13 which gives the relationship between the trail length L, drag radius r, fuel consumption per cm of path $\varphi$, and amount of water which is condensed as the result of combustion of each gram of fuel $\lambda$, shows that for the piston type aircraft,

$$L = \frac{\text{const.}}{r}\left(\frac{\varphi\lambda}{Ur}\right)^{1+m} \qquad 6.09$$

so that in the case of jet aircraft the relationship becomes

$$L = \frac{\text{const. } U}{C^2 V}\left(\frac{\varphi\lambda}{Ur}\right)^{1+m} \qquad 6.10$$

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London), M.R.P. No. 831 by Mr. M.J. Blackwell. A copy is available in the Library of the Meteorological Office."

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S.C.III/162
3rd September, 1953.

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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

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A standard eye has been specified by the International Commission on Illumination from the results of experiments on numerous observers (3). This specification states the relative sensitivity of the eye, throughout the spectrum, to a given quantity of monochromatic radiation. Although the sensitivity function varies with intensity at lower levels of illumination, it is constant in the range with which this report is concerned (photopic vision). This puts the comparison of illumination due to sources of different colours on a sound basis, and facilitates instrumental rather than visual photometry, it being merely necessary to ensure that the relative response of the intrument is the same as that of the I.C.I. eye throughout the spectrum.

The unit of illumination is the lux, the illumination produced by a source of unit candle power (1 candela) at a distance of 1 metre, where 1 candela is one sixtieth of the luminous intensity per sq. cm. of the standard radiator. For daylight illumination, the kilolux is a convenient unit. In the present work, mean hourly values in kilolux have been tabulated and totals of these have been used. The tabulated values are therefor in kilolux-hours, a convenient rather than physically significant unit.

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(b) History of the Kew recorder. The recording of daylight illumination at Kew Observatory was planned by Dr. J.M. Stagg in 1939, and the project was revived in 1946. Continuous records had been maintained at Plymouth for some years by Atkins (4) using a vacuum sodium cell, and experiments in continuous recording at the National Physical Laboratory (5) had involved the use of a selenium cell. In deciding on the arrangement to be used at Kew, valuable advice was given by Dr. J.S. Preston of the N.P.L. and the final choice was that of a selenium photocell combined with a specially constructed correcting filter, exposed below a diffusing surface of pot opal glass which was roughened on the upper side. The filter was designed to match the mean response curve of the class of cell used, to allow replacement of the cell with the least possible loss of precision.

For the first two years, the instrument which will be referred to as the Mk.I was in use; this was then replaced by that known as the Mk.II. The change was made chiefly for convenience in handling and maintenance, but it was later realised that systematic errors of calibration had been reduced by the change.

As experience was gained, appreciation of errors in the method of measurement and calibration increased; but most of these were not fully apparent until an examination of the first five years of recording had been made. It was found possible to correct the tabulated mean values retrospectively, and much of this report is taken up with a discussion of the imperfections of the measurements and of the corrections which were applied.

## 2. Description of the instruments.

(a) Mark I instrument. This, the original version of the recorder, was used throughout 1947-48. As shown in Fig. 1(a), the cell and filter unit was mounted within an airtight metal box with a glass cover. This box was set inside a louvred screen, at the top of which was a sheet of diffusing pot opal glass, roughened on the upper side. This, in turn, was protected by a clear, hemispherical glass dome. The air within the box containing the cell was kept dry by silica-gel which was changed once a month at times of calibration.

(b) Mark II instrument. After 1948, the more compact Mk.II version came into use. As shown in Fig. 1(b), the pot opal glass now directly surmounts the inner box which contains the cell and filter unit. The hemispherical glass dome is mounted as an integral part of a new outer case which, in turn, is covered by the white louvred screen.

(c) Recorder circuit. The current output against intensity of illumination curve, for the type of photocell used, depends on the load resistance. To obtain a nearly linear relation, this resistance is best fixed at about 200 ohms. The output from the cell is led away to a Cambridge thread-recorder which is combined with a network of shunt and series resistors, arranged in such a way as to allow three sensitivities (for different times of the year) while keeping the load resistance constant at 200 ohms.

(d) Protection of the photocell from excessive exposure. Since excessive illumination, and consequent high current output, would impair the performance of the photocell, the geometry of the Mk.I instrument was designed to reduce the incident daylight flux by a factor of about ten. This is achieved partly by means of the dimensions and distances of the apertures and partly by the losses due to reflection at the hemispherical glass dome, the diffusing plate, the inner glass cover and the filter.

In the case of the Mk. II instrument, the photocell is rather closer to the diffusing plate, and an additional restricting diaphragm is inserted underneath and touching the plate, in order to limit the aperture. In this way, the maximum permissible output current is not reached under working conditions.

(e) Response curves of instrument and standard eye. Fig. 2 shows the close approximation of the sprectral sensitivity curve of the cell and filter, as determined at the N.P.L., to that of the I.C.I. standard eye, which represents the normal human eye adapted to conditions of high luminance (photopic vision).

## 3. Standardisation of the recorder.

(a) Method. The difficulty is that it is not possible to produce a known illumination comparable in intensity with natural daylight. It is easy, however, to produce, on the cell and filter unit alone, a known illumination comparable with that to which it is subjected when the diffusing plate is exposed to full daylight, and the unit is in its normal position inside the recorder.

The combined cell and filter unit is therefore mounted vertically at one end of a photometry bench. A 100 watt lamp is used to obtain a response curve at various illuminations up to the maximum intensity expected, intensity being judged by cell output. Any marked non-linearity in the response curve, (output being plotted against the inverse square of distance), leads to replacement of the photocell.

In the second stage of calibration, the complete recorder, again mounted vertically, is exposed to a 1000 watt standard lamp at one or two points on the photometry bench, and the cell output noted. In this way, the scale of illumination for the first curve is computed in absolute units (kilolux). Allowance is made for any stray light by alternately shading and unshading at a point midway between the lamp and the pot opal disc of the recorder.

According to the maximum intensity expected during the month to which the calibration applies, the best straight line is drawn through the experimental points, within this range, on the first curve. The factor is then calculated in kilolux per division of the thread-recorder chart, for whichever of the three sensitivities is to be used.

(b) Standard lamps. Two 1000 watt, 110 volt, tungsten filament projector lamps, calibrated at the National Physical Laboratory in 1947, and recalibrated in 1952, have been used as the reference sources throughout the period. As the two calibrations agreed to within 2% and the quoted accuracy is given as ±2%, these standards are taken to have remained constant in brightness. The lamps are slightly under-run at 100 volts, when the horizontal candle power is about 2200 C.P.

## 4. Systematic errors of Standardisation.

(i) From the smoothness of the calibration curves, and the fact that these pass through the zero point, it is concluded that errors of measurement, and errors resulting from stray light and rapid uncompensated voltage fluctuations may be neglected. Allowance must be made, however, for the polar diagram of the lamps and the variation of illumination over the area of the opal plate. In the case of the Mk. I instrument, the diameter of this plate is six inches, involving a difference of 6% between the mean intensity and that at the centre of the opal disc. In the Mk. II instrument, the limiting diaphragm has a diameter of two inches, involving a correction of -1%.

(ii) There is also a small systematic correction for the difference in quality of the calibrating light and daylight, since the combined cell plus filter response curve is not exactly that of the I.C.I. eye. The filter matches the photocell correctly for light of colour temperature 6000°K. For the 2500°K of the tungsten filament lamps, the matching is imperfect and necessitates a further correction of -2%. This was computed from the data shown in Fig. 3, for which the energy distribution of extra-terrestrial solar radiation is taken from Nicolet (1951).

(iii) A further systematic error in the calibrations, made during the period under discussion, is due to incorrect measurement of the lamp voltage. By an oversight the voltage was measured across the supply leads instead of at the lamp filament, as specified in the N.P.L. standardisation. Fortunately the same leads had been used throughout the period, and a correction of 5% was determined experimentally and applied to all tabulated results.

5. Systematic errors of recording. The corrections considered above are fixed for the particular instruments and the methods of calibration used. The following corrections, though based on specific instrumental and physical factors, are computed from knowledge of the particular local conditions under which the instruments have been exposed, and on the weather at Kew Observatory during the period 1947-51. It would appear permissible, however, to regard the corrected data as representative of similar sites to an accuracy of the order of ±5%, for the same period.

(a) Exposure of the recorder. The instrument is exposed on the south roof of the Observatory, at a point where some 10% of the north sky is obstructed by the Observatory dome. It is calculated that the fraction of the luminous flux from the sky, which is thus prevented from falling on the horizontal opal plate, is only 3%, or on average some 1-2% of the total flux. On many occasions when the sun is shining, back reflection from the dome contributes a compensating 1-2% of the total luminous flux on the instrument. It is therefore concluded that errors from imperfect exposure are only of the order of 1%.

(b) Temperature coefficient of the thread-recorder. Each time the instrument is calibrated, the sensitivity of the thread-recorder galvanometer is also found, and its temperature noted. Subsequent variations of room temperature, in the basement in which it is situated, may cause random errors of 1-2% due to the temperature coefficient of the galvanometer, but any residual systematic errors in mean values of daylight may be neglected.

(c) Temperature coefficient of the photocell. This varies, for the class of cell used, with the external circuit resistance and with the intensity of illumination. With typical working conditions simulated in the laboratory, the coefficient of the cell and filter was found to be -0.3% per 1°C between 0° and 60°C.

The temperature of the cell had not been noted, either during calibration or recording. As a first step to correcting the tabulated mean values, the excess of cell temperature over its surroundings was measured (by thermocouples) during a typical calibration, and during normal recording in various conditions of sky, wind and time of day. Values of room temperature were then extracted for all available calibration records, and cell temperatures pertaining to each calibration during the five year period were estimated.

From the experimental data obtained during normal recording, it was possible to relate the excess temperature of the cell over its surroundings to the sine of the sun's elevation for cloudless periods, and to equate the cell temperature to that of the surroundings (i.e. screen temperature) for overcast periods. Having drawn diurnal curves for typical cloudless days in each of the four seasons by fixing the maximum excess temperature at 1300 Z and completing the curves with reference to the sun's elevation, and using mean diurnal curves of screen temperature for the overcast days, interpolation could then be made for other cloud conditions.

For the immediate purpose of applying these corrections to hourly means for 10-day periods, the average cloud amount for Kew during daylight hours was found to be 5/8. The interpolated curves of cell temperature for 5/8 cloud were finally compared with the curve of calibration temperatures, and the differences between them, when multiplied by the temperature coefficient, then gave the corrections to apply. These range between -3% and +5% according to time of day and year.

(d) Imperfection of the diffusing plate - deviation from the cosine law. The response of the daylight recorder, to light incident at various angles, depends largely on the nature of the roughened upper surface of the pot opal glass, but partly also on the geometrical layout of the cell, apertures and diaphragm. The response, found by experiment, is represented in Fig. 4 as a percentage of the response given by a perfect receiver, i.e: one conforming to the cosine law. It does not differ significantly between the Mk.I and Mk.II instruments. The errors at grazing incidence amount to some 50%, and lengthy computations were necessary before the basic tabulations could be corrected.

The original tabulations were first analysed into hourly means (for the five years 1947-51) for the first, second and last 10-day period in each month, (i.e: 1st to 10th, 11th to 20th, and 21st to end of month), designated by the headings 0, 1, 2. Each of these hourly means was split into the direct and diffuse components of daylight by assuming, for the purpose in hand, that the relative proportion of these components is the same as for direct and diffuse solar radiation. These latter quantities were available for the same period, being part of the programme of measurements of short wave solar radiation at Kew Observatory.

The relative intensity of diffuse short wave radiation, over the hemisphere, has been investigated for cloudless and overcast skies by Dines and Dines (1927), whose results do not differ significantly, for the present purpose, from the distribution function quoted by Middleton (1952). By graphical integration of the response of the recorder in zones over the whole hemisphere, a total response of 89.7% is obtained for overcast days and 86.3% for cloudless days. Interpolating for the 5/8 cloud of average conditions gives a mean response of 88.6%, involving a correction of 11.4% to the diffuse component of daylight.

Lastly, the mean elevation of the sun was computed for each hour of each 10-day period. The corresponding percentage correction was obtained from the response curve and applied to the direct component of daylight, and recombined with the diffuse component.

6. The corrected hourly means. The total percentage correction for the separate errors (temperature coefficient, cosine law and calibration errors) was finally computed. Table I shows the complete correction table which has been used. It will be seen that, apart from some large percentage corrections to very small hourly means near sunrise and sunset, the figures range between +1% and +10%.

The corrected hourly means for 10-day periods are presented in Table II. The basic tabulations, from which these means are extracted, are obtained by estimating the mean intensity for the hours in question by placing a graduated transparent scale over the thread-recorder charts. Random errors up to ±10% may, on occasion, be introduced by this procedure, especially on days of broken cloud and rapidly varying intensity. [*] More generally these errors are around ±5% [*], and any residual systematic errors in the means (each based on 50 observations) will be of the order of ±1%. The times given are in G.M.T.

7. The variability of daylight. Provisional uncorrected data, based on the first two years of recording with the Mk.I instrument, have been given by Stagg (1949) in a paper which had a restricted circulation. These may be related to the present data, when all occasions are considered together, by means of the Table I corrections reduced by 3% throughout (i.e. apply +4% for +7%, etc).

The intensity of daylight illumination on a horizontal surface above the earth's atmosphere, depends on the time of day and year, and on the latitude of the observing station. At Kew (latitude 51° 28'N), the intensity attains a value of 121 kilolux at noon, in midsummer. At noon in midwinter, this value falls to only 37 kilolux.

On cloudless days, these extraterrestrial intensities are not exceeded at the ground, but, with small amounts of cloud near but not obstructing the sun, the hourly mean can rise to about 130 kilolux, owing to scattering and reflection at the cloud surfaces. A value of 200 kilolux may be reached for a minute or two, particularly when large Cumulus clouds are present, but this only rarely.

To demonstrate the longer term variability of daylight illumination, the annual totals for the years 1947 to 1951 have been extracted. These corrected totals, in megalux-hours, are compared with values for the years 1947 to 1949 obtained at Plymouth by Atkins and Jenkins (1952), who used a vacuum sodium cell.

| | 1947 | 1948 | 1949 | 1950 | 1951 |
|---|---|---|---|---|---|
| Kew | 118 | 125 | 114.5 | 111 | 109 |
| Plymouth | 108 | 124 | 121 | - | - |

The fluctuations of successive 10-day means, for the same hours in Table II, also demonstrate the considerable variations possible, even when five years data are included.

8. Diurnal variations of daylight. In Fig. 5, mean diurnal curves are shown for the months of March, June, September and December, the time-axis having been converted to L.A.T. Also included is a composite curve for the Equinoxes, which has been constructed from two separate curves, based on the values for the 10-day periods Mar. 2 and Sept. 2.

Fig. 6 shows the mean diurnal curves for selected cloudless days, for the same four months. To construct these, the original charts were re-examined and all occasions selected where there was 1/8 cloud or less (and this away from the immediate vicinity of the sun) for at least three consecutive hours. The end hours of these periods were rejected and the remaining hours tabulated under time of day. A correction table, similar in form to Table I, was then constructed and applied to the means for each hour.

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[*] Footnote: These estimates, based on trials in which many charts were measured independently by different observers, are due to Mr. R.H. Eldridge.

The same procedure was carried out for overcast days; periods being selected when there were 7/8 cloud or more (the sun itself always obscured) for at least three hours. When the middle hours were tabulated under time of day and the corresponding correction table applied, it became apparent that insufficient data existed for mean diurnal curves to be drawn objectively. The results are therefore included in Table III with the number of occasions contributing to each hourly mean given in parenthesis.

9. Frequency distribution of daylight intensity. The basic tabulations for the four months: March, June, September and December, were reanalysed for the midday hours: (11-12) and (12-13). For each month, the range of hourly values was subdivided into three equal parts. To the lowest range were applied the corrections for overcast days; to the middle range, those of Table I for average days; to the highest range, those for cloudless days. Slight adjustments were necessary to smooth the corrections to hourly values on the borderline between two ranges.

The corrected tabulations were grouped in ranges of 15 kilolux for the midday hours of March, June and September; 10 kilolux ranges being used for December because of the concentration of these observations into the lowest ranges. Histograms representing the frequency distribution of intensiites, thus found, are shown in Fig. 7.

10. Luminous efficiency of daylight. While daylight is being recorded in the manner outlined above, a nearby Moll solarimeter is simultaneously recording "total" solar radiation in the waveband 0.3μ to about 3.0μ , as described by Stagg (1950). When the ratio, daylight intensity: radiation intensity is expressed in the form, lumens per watt, it is spoken of as the "luminous efficiency of daylight".

To form estimates of this quantity, values of radiation were extracted for the same occasions with cloudless skies as were considered in section 8. It was found unnecessary to correct these tabulations, since: the systematic errors, present in the calibrations of the daylight recorder, are not encountered in the calibrations of the Moll solarimeter (in situ, against the Ångström pyrheliometer), and the temperature coefficient and cosine law corrections for this particular Moll solarimeter are very nearly equal, and of opposite sign.

The results are given in Table IV, where K stands for values of daylight intensity in kilolux and T for solar radiation intensity in mw/cm$^2$. The main results may be summarised by giving the mean luminous efficiency for each of the four months (for cloudless days only):

| | Mar. | Jun. | Sept. | Dec. | Year |
|---|---|---|---|---|---|
| lumens/watt<br>(limits) | 119.5<br>(95-132) | 130.0<br>(121-144) | 137.5<br>(107-152) | 112.5<br>(103-149) | 128.5<br>(95-152) |

There is no reason to suspect these values, for it will be shown in a later paper that the illumination and radiation tabulations satisfy several checks of mutual consistency, but they are higher than those usually quoted. Kimball (1924) found values between 97.5 (solar elevation 22°) and 108.0 (solar elevation 65°), while Atkins and Poole (1936) found 107 lumens/watt when some cloud was also present. On the basis of Nicolet's (1951) data for the spectral distribution of extra-terrestrial solar energy, the luminous efficiency of extra-terrestrial daylight is computed to be 104 lumens/watt. With high water vapour content in the atmosphere, (made effectively still higher for large values of air mass), it is to be expected that infra-red absorption would, in causing depletion of total radiation, lead to greater luminous efficiency.

This is to some extent supported by Fig. 8 in which the mean efficiencies for the four months have been plotted against the mean water vapour contents for the whole atmosphere. Other features noted in the individual values are the lack of variation of the means with solar elevation but the much greater scatter for low angles. This suggests that the variations of the degree of atmospheric pollution, on otherwise similar cloudless days, have a large effect on luminous efficiency.

11. Conclusion and suggestions for further work. As far as the preceding data on daylight illumination can be applied to similar sites, it is thought that the corrected means are within ±5% of the true values. The limited data available for selected overcast and cloudless days are more subject to sampling errors although the qualitative relations can be regarded as fairly representative. In particular, the nearness of the site to the smoke pollution source of the Greater London area often affects the data for cloudless (and frequently calm) days.

The instruments and recording details can clearly be improved. Among projected modifications can be mentioned the installation of a smaller diffusing plate more in keeping with the cosine law, and the thermostatic control of the photocell temperature.

12. Acknowledgments. I wish to express my thanks to Dr. G.D. Robinson for his guidance throughout the investigation, and in particular for checking the mutual consistency of the daylight illumination measurements with those of the other short wave solar radiation components which are measured at Kew Observatory. Thanks are also due to Mr. D.B.B. Powell for assistance in all stages of the breakdown of the data and investigation of instrumental errors, and to Mr. J. MacDowall for help in estimating certain errors in calibration and of exposure.

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## Table I.

Percentage correction added to original 5-year mean Daylight Tabulations

| | | 4 - 5 | 5 - 6 | 6 - 7 | 7 - 8 | 8 - 9 | 9 - 10 | 10-11 | 11-12 | 12-13 | 13-14 | 14-15 | 15-16 | 16-17 | 17-18 | 18-19 | 19-20 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| J | 0 | | | | | 6 | 4 | 4 | 5 | 5 | 4 | 4 | 6 | | | | |
| | 1 | | | | | 5 | 6 | 6 | 7 | 6 | 6 | 6 | 4 | | | | |
| | 2 | | | | | 4 | 3 | 6 | 5 | 4 | 6 | 6 | 5 | | | | |
| F | 0 | | | | 10 | 6 | 6 | 5 | 5 | 7 | 6 | 8 | 8 | 10 | | | |
| | 1 | | | | 10 | 10 | 5 | 6 | 5 | 7 | 7 | 6 | 9 | 11 | | | |
| | 2 | | | | 6 | 6 | 4 | 4 | 5 | 5 | 5 | 5 | 7 | 6 | | | |
| M | 0 | | | 8 | 7 | 7 | 4 | 5 | 6 | 5 | 7 | 5 | 8 | 6 | 1 | | |
| | 1 | | | 10 | 5 | 5 | 4 | 4 | 5 | 5 | 5 | 6 | 6 | 7 | 9 | | |
| | 2 | | | 8 | 6 | 6 | 3 | 4 | 3 | 3 | 5 | 4 | 6 | 8 | 7 | | |
| A | 0 | | 10 | 10 | 7 | 6 | 4 | 3 | 4 | 4 | 4 | 5 | 6 | 7 | 10 | 2 | |
| | 1 | | 9 | 8 | 7 | 4 | 3 | 5 | 3 | 3 | 5 | 4 | 6 | 8 | 8 | 10 | |
| | 2 | | 11 | 8 | 6 | 5 | 4 | 2 | 3 | 3 | 4 | 5 | 8 | 7 | 10 | 9 | |
| M | 0 | 16 | 9 | 8 | 7 | 6 | 6 | 4 | 3 | 3 | 5 | 7 | 8 | 9 | 12 | 10 | 3 |
| | 1 | 10 | 8 | 7 | 7 | 4 | 3 | 3 | 3 | 3 | 5 | 5 | 7 | 9 | 10 | 10 | 3 |
| | 2 | 10 | 7 | 7 | 5 | 5 | 3 | 2 | 3 | 3 | 4 | 5 | 6 | 7 | 10 | 8 | 10 |
| J | 0 | 10 | 9 | 8 | 4 | 4 | 2 | 2 | 3 | 3 | 3 | 4 | 6 | 7 | 10 | 9 | 10 |
| | 1 | 8 | 7 | 8 | 4 | 4 | 3 | 2 | 3 | 3 | 3 | 4 | 7 | 7 | 9 | 9 | 10 |
| | 2 | 11 | 7 | 7 | 4 | 3 | 3 | 2 | 3 | 3 | 3 | 4 | 7 | 7 | 9 | 10 | 10 |
| J | 0 | 6 | 6 | 6 | 4 | 1 | 2 | 1 | 2 | 2 | 2 | 3 | 5 | 5 | 9 | 9 | 8 |
| | 1 | 9 | 6 | 7 | 4 | 3 | 1 | 1 | 2 | 3 | 3 | 4 | 5 | 6 | 7 | 6 | 8 |
| | 2 | 10 | 8 | 6 | 6 | 2 | 1 | 1 | 2 | 2 | 2 | 3 | 5 | 8 | 8 | 10 | 4 |
| A | 0 | 0 | 7 | 6 | 5 | 4 | 2 | 2 | 2 | 2 | 4 | 2 | 6 | 7 | 9 | 9 | 1 |
| | 1 | | 8 | 9 | 6 | 4 | 3 | 2 | 3 | 3 | 3 | 4 | 6 | 7 | 10 | 9 | |
| | 2 | | 9 | 7 | 7 | 4 | 3 | 2 | 2 | 2 | 3 | 4 | 6 | 9 | 9 | 10 | |
| S | 0 | | 13 | 9 | 6 | 6 | 6 | 3 | 2 | 2 | 5 | 6 | 7 | 7 | 9 | 12 | |
| | 1 | | 0 | 5 | 8 | 6 | 5 | 3 | 5 | 5 | 4 | 6 | 8 | 9 | 8 | | |
| | 2 | | | 7 | 8 | 7 | 7 | 5 | 5 | 5 | 6 | 8 | 9 | 9 | 10 | | |
| O | 0 | | | 14 | 7 | 6 | 5 | 4 | 5 | 5 | 6 | 7 | 8 | 9 | 2 | | |
| | 1 | | | | 7 | 8 | 7 | 5 | 7 | 6 | 6 | 7 | 9 | 8 | | | |
| | 2 | | | | 6 | 6 | 5 | 5 | 5 | 5 | 6 | 9 | 9 | 16 | | | |
| N | 0 | | | | 8 | 9 | 9 | 7 | 8 | 8 | 7 | 9 | 7 | 9 | | | |
| | 1 | | | | 16 | 6 | 5 | 6 | 6 | 6 | 8 | 6 | 5 | | | | |
| | 2 | | | | | 9 | 6 | 8 | 6 | 7 | 8 | 8 | 7 | | | | |
| D | 0 | | | | | 7 | 4 | 5 | 6 | 7 | 6 | 5 | 7 | | | | |
| | 1 | | | | | 2 | 3 | 5 | 6 | 6 | 5 | 6 | 6 | | | | |
| | 2 | | | | | 14 | 5 | 5 | 6 | 6 | 6 | 4 | 8 | | | | |

## Table II.

Corrected hourly means of Daylight Illumination (kilolux-hours) for 10-day periods.

| | | 4 - 5 | 5 - 6 | 6 - 7 | 7 - 8 | 8 - 9 | 9 - 10 | 10-11 | 11-12 | 12-13 | 13-14 | 14-15 | 15-16 | 16-17 | 17-18 | 18-19 | 19-20 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| J | 0 | | | | | 1.6 | 5.6 | 9.7 | 12.1 | 12.2 | 9.1 | 4.9 | 1.7 | Tr | | | |
| | 1 | | | | | 1.9 | 7.1 | 11.6 | 15.1 | 13.6 | 10.8 | 6.9 | 2.8 | Tr | | | |
| | 2 | | | | Tr | 2.3 | 6.4 | 11.4 | 14.9 | 14.2 | 11.7 | 8.1 | 3.9 | 0.5 | | | |
| F | 0 | | | | 0.3 | 3.2 | 8.3 | 13.2 | 14.9 | 16.5 | 15.4 | 10.1 | 5.1 | 1.3 | Tr | | |
| | 1 | | | | 1.0 | 5.9 | 11.1 | 18.9 | 20.8 | 20.3 | 18.2 | 13.7 | 8.5 | 3.0 | Tr | | |
| | 2 | | | Tr | 2.4 | 8.7 | 16.6 | 23.7 | 23.6 | 26.1 | 22.8 | 19.7 | 12.7 | 4.7 | 0.4 | | |
| M | 0 | | | 0.2 | 3.9 | 9.9 | 16.0 | 22.8 | 27.7 | 27.6 | 25.2 | 18.8 | 13.3 | 5.8 | 1.0 | | |
| | 1 | | | 1.6 | 7.1 | 15.0 | 23.0 | 28.7 | 30.0 | 33.2 | 26.9 | 23.6 | 17.0 | 9.4 | 2.4 | Tr | |
| | 2 | | Tr | 3.3 | 11.6 | 21.0 | 27.1 | 34.0 | 36.0 | 37.3 | 33.6 | 28.5 | 20.8 | 12.3 | 3.7 | Tr | |
| A | 0 | | 0.8 | 7.6 | 19.8 | 33.2 | 38.4 | 42.7 | 50.0 | 47.3 | 42.4 | 31.8 | 26.9 | 16.8 | 7.2 | 0.7 | |
| | 1 | | 2.3 | 11.0 | 23.4 | 35.9 | 45.6 | 55.9 | 59.2 | 60.3 | 55.4 | 44.3 | 35.7 | 22.4 | 9.9 | 2.4 | |
| | 2 | Tr | 4.1 | 13.3 | 27.6 | 40.4 | 52.6 | 58.3 | 62.3 | 58.7 | 57.1 | 48.8 | 36.7 | 25.0 | 11.2 | 2.9 | |
| M | 0 | 0.4 | 4.7 | 13.1 | 22.3 | 30.2 | 41.2 | 50.9 | 52.0 | 50.6 | 50.4 | 45.9 | 37.6 | 28.3 | 15.7 | 5.3 | 0.1 |
| | 1 | 1.0 | 7.7 | 18.2 | 30.1 | 41.7 | 50.5 | 61.9 | 62.1 | 57.7 | 58.0 | 48.1 | 37.2 | 26.6 | 17.5 | 5.7 | 0.5 |
| | 2 | 1.8 | 7.5 | 16.8 | 28.9 | 40.7 | 47.4 | 54.2 | 56.4 | 59.6 | 55.7 | 48.8 | 40.7 | 28.5 | 18.5 | 7.3 | 1.7 |
| J | 0 | 3.4 | 13.8 | 27.4 | 38.5 | 52.5 | 62.7 | 73.8 | 75.7 | 79.1 | 71.6 | 64.0 | 52.0 | 37.9 | 24.6 | 10.3 | 2.7 |
| | 1 | 3.4 | 11.1 | 25.7 | 37.2 | 50.5 | 56.9 | 63.4 | 63.8 | 63.6 | 63.4 | 58.8 | 49.6 | 39.2 | 24.0 | 11.5 | 3.6 |
| | 2 | 3.1 | 11.7 | 22.9 | 34.9 | 48.6 | 57.6 | 64.0 | 66.4 | 67.5 | 65.7 | 63.8 | 50.1 | 38.2 | 23.1 | 11.1 | 3.1 |
| J | 0 | 2.7 | 10.7 | 21.5 | 37.2 | 45.0 | 58.1 | 61.2 | 65.7 | 63.7 | 60.8 | 52.8 | 43.8 | 31.3 | 21.9 | 10.7 | 3.2 |
| | 1 | 1.8 | 8.9 | 20.0 | 36.3 | 42.5 | 54.3 | 61.9 | 71.0 | 64.0 | 59.2 | 52.3 | 42.2 | 30.6 | 21.2 | 8.9 | 2.5 |
| | 2 | 1.3 | 8.8 | 21.0 | 37.5 | 46.6 | 58.1 | 67.4 | 71.6 | 72.4 | 69.9 | 58.7 | 46.0 | 34.8 | 22.9 | 10.0 | 2.3 |
| A | 0 | Tr | 5.9 | 16.9 | 28.4 | 40.6 | 49.0 | 57.5 | 63.0 | 61.8 | 63.0 | 52.6 | 42.6 | 30.4 | 19.0 | 8.3 | 0.7 |
| | 1 | Tr | 4.7 | 16.2 | 30.7 | 45.1 | 57.2 | 65.5 | 68.0 | 68.9 | 63.7 | 53.4 | 43.3 | 29.2 | 16.5 | 5.6 | 0.1 |
| | 2 | | 2.5 | 11.8 | 26.0 | 41.1 | 53.8 | 58.0 | 63.5 | 62.6 | 59.9 | 47.0 | 37.4 | 26.6 | 13.5 | 3.1 | |
| S | 0 | | 1.0 | 8.6 | 21.2 | 35.8 | 47.4 | 55.4 | 61.4 | 56.4 | 53.9 | 44.2 | 33.7 | 17.7 | 7.6 | 0.8 | |
| | 1 | | 0.3 | 6.4 | 18.1 | 30.9 | 39.8 | 46.5 | 47.7 | 47.5 | 43.9 | 36.7 | 27.5 | 15.4 | 5.2 | Tr | |
| | 2 | | Tr | 4.0 | 13.1 | 25.0 | 34.2 | 41.7 | 45.1 | 44.3 | 39.2 | 34.0 | 23.0 | 10.1 | 2.5 | | |
| O | 0 | | | 1.7 | 7.5 | 17.2 | 26.7 | 36.6 | 39.3 | 39.3 | 37.5 | 26.1 | 15.2 | 6.0 | 0.4 | | |
| | 1 | | | 0.9 | 6.0 | 14.3 | 23.7 | 26.8 | 33.2 | 31.0 | 26.3 | 17.8 | 10.9 | 3.3 | Tr | | |
| | 2 | | | Tr | 3.2 | 9.2 | 14.4 | 20.5 | 24.1 | 24.1 | 19.6 | 13.9 | 6.6 | 1.2 | | | |
| N | 0 | | | Tr | 2.2 | 8.3 | 14.8 | 20.3 | 22.8 | 21.5 | 17.6 | 10.7 | 4.4 | 0.2 | | | |
| | 1 | | | | 1.1 | 6.1 | 12.4 | 18.0 | 19.3 | 18.7 | 14.7 | 8.5 | 2.9 | Tr | | | |
| | 2 | | | | 0.2 | 3.8 | 8.0 | 13.0 | 15.6 | 14.6 | 10.9 | 5.5 | 1.7 | Tr | | | |
| D | 0 | | | | Tr | 2.2 | 6.4 | 9.9 | 11.2 | 12.0 | 8.9 | 5.0 | 1.3 | Tr | | | |
| | 1 | | | | | 1.6 | 5.5 | 9.6 | 12.1 | 11.1 | 8.0 | 4.9 | 1.2 | Tr | | | |
| | 2 | | | | | 1.4 | 4.4 | 7.9 | 10.1 | 10.6 | 7.8 | 4.4 | 1.2 | Tr | | | |

## Table III.

Diurnal variation of Daylight Illumination (kilolux-hours) on overcast days

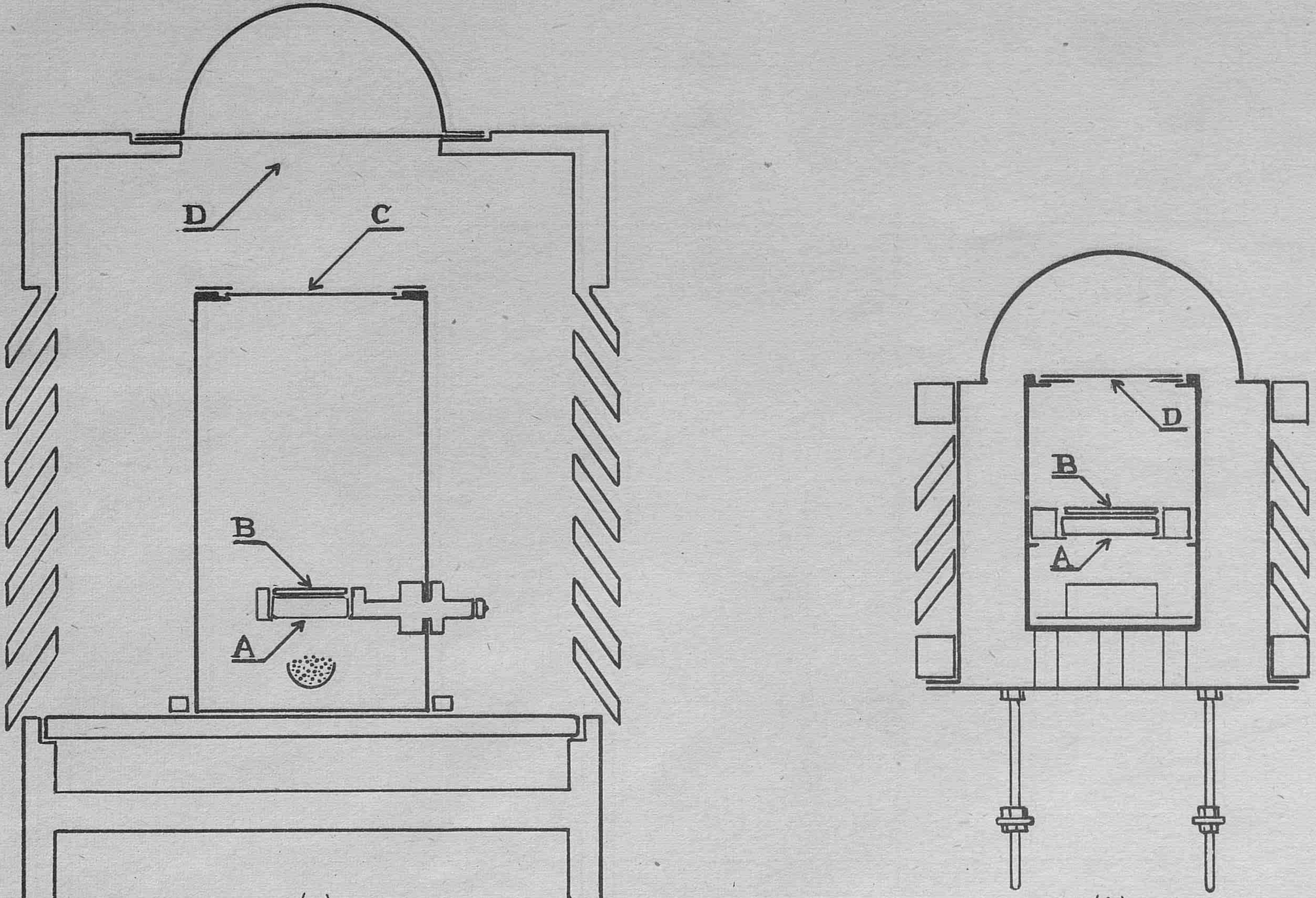
| | | 5 - 6 | 6 - 7 | 7 - 8 | 8 - 9 | 9 - 10 | 10-11 | 11-12 | 12-13 | 13-14 | 14-15 | 15-16 | 16-17 | 17-18 | 18-19 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mar. | | | - | - | 8½ | 6½ | 7½ | 10 | 9 | 9 | 9 | 11 | 6 | - | |
| | | | - | - | (2) | (1) | (1) | (14) | (14) | (17) | (19) | (9) | (1) | - | |
| June | | - | 8 | 13 | - | - | - | 11 | 15 | 20½ | 11 | 12 | 8 | 4½ | - |
| | | - | (10) | (7) | - | - | - | (1) | (1) | (1) | (1) | (2) | (3) | (2) | - |
| Sep. | | | - | 9 | - | - | 22 | 17½ | 16 | 17½ | 15 | 12 | 9 | - | |
| | | | - | (5) | - | - | (2) | (10) | (9) | (8) | (12) | (9) | (1) | - | |
| Dec. | | | | | - | - | - | 5½ | 6 | 5 | - | - | | | |
| | | | | | - | - | - | (17) | (22) | (21) | - | - | | | |

## Table IV.

Luminous efficiency of Daylight (lumens/watt) on cloudless days

| | | 5 - 6 | 6 - 7 | 7 - 8 | 8 - 9 | 9 - 10 | 10-11 | 11-12 | 12-13 | 13-14 | 14-15 | 15-16 | 16-17 | 17-18 | 18-19 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Mar. | K | | - | 15 | 26 | - | - | 65 | 64 | 61 | 46 | 36 | 20 | - | |
| | T | | - | 15½ | 27½ | - | - | 51 | 52½ | 47½ | 39½ | 27½ | 18 | - | |
| | K/T | | - | 96 | 95 | - | - | 127 | 122 | 128 | 116 | 132 | 111 | - | |
| June | K | 21 | 31 | 51 | 74 | 86 | 48 | 103 | 105 | 98 | 86 | 72 | 58 | 41 | 20 |
| | T | 15 | 25½ | 40 | 56 | 67½ | 75½ | 82 | 82½ | 74½ | 65 | 56 | 43 | 28½ | 15½ |
| | K/T | 139 | 121 | 125 | 133 | 128 | 130 | 125 | 127 | 132 | 132 | 129 | 135 | 144 | 128 |
| Sep. | K | | - | 24 | 41 | - | - | 90 | 90 | 76 | 55 | 36 | - | - | |
| | T | | - | 20 | 35 | - | - | 61½ | 60 | 50 | 40 | 33½ | - | - | |
| | K/T | | - | 121 | 118 | - | - | 146 | 150 | 152 | 137 | 107 | - | - | |
| Dec. | K | | | | - | - | 20 | - | 20 | 15 | 13 | - | | | |
| | T | | | | - | - | 18 | - | 19 | 14½ | 8½ | - | | | |
| | K/T | | | | - | - | 111 | - | 104 | 103 | 149 | - | | | |

1/4 Scale



(a)

Mark I instrument

(b)

Mark II instrument

Fig.1

A. Selenium photocell.
B. Filter.
C. Clear glass cover.
D. Opal diffusing glass.

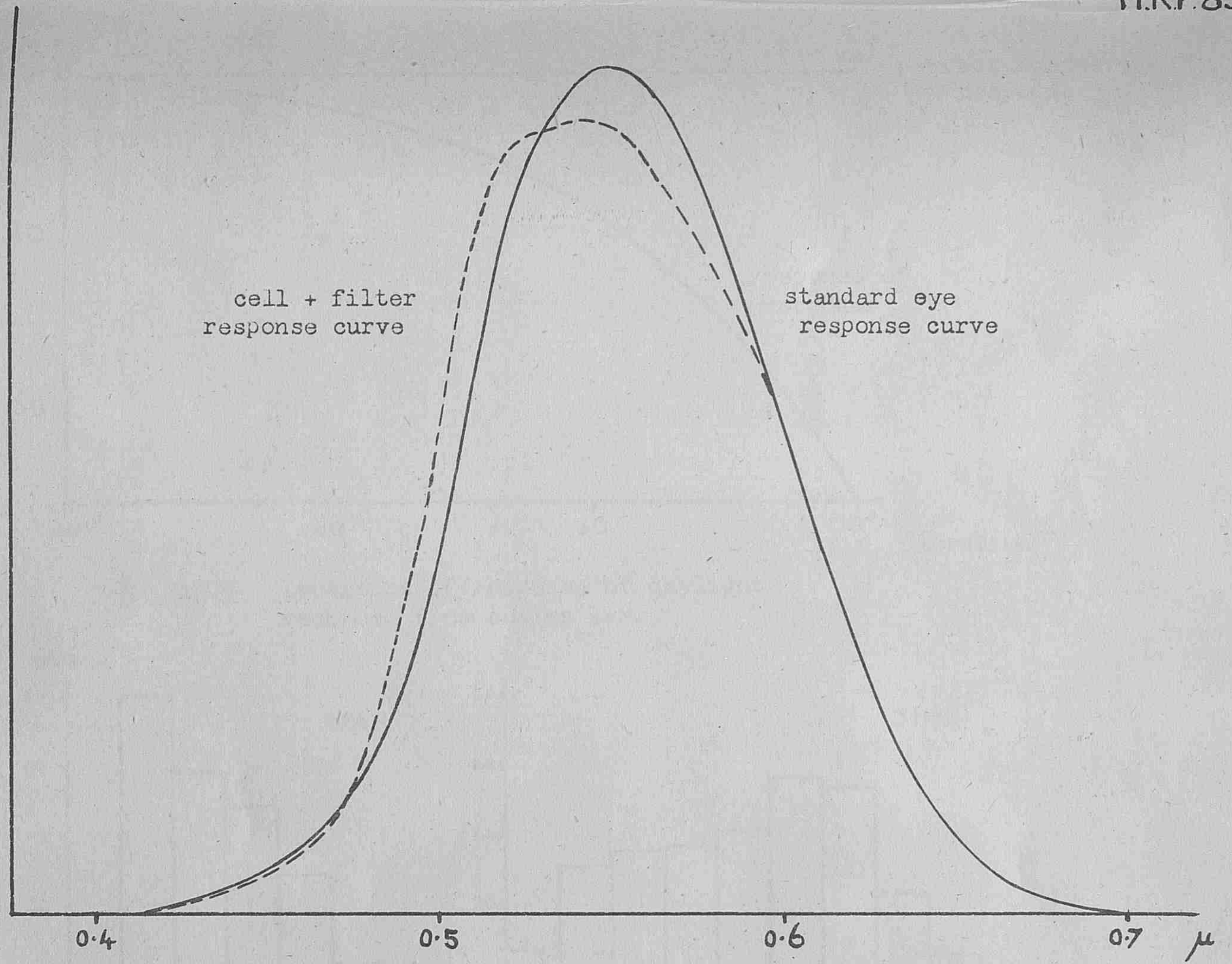


Fig.2 - Relative spectral sensitivity curves of cell + filter and standard eye.

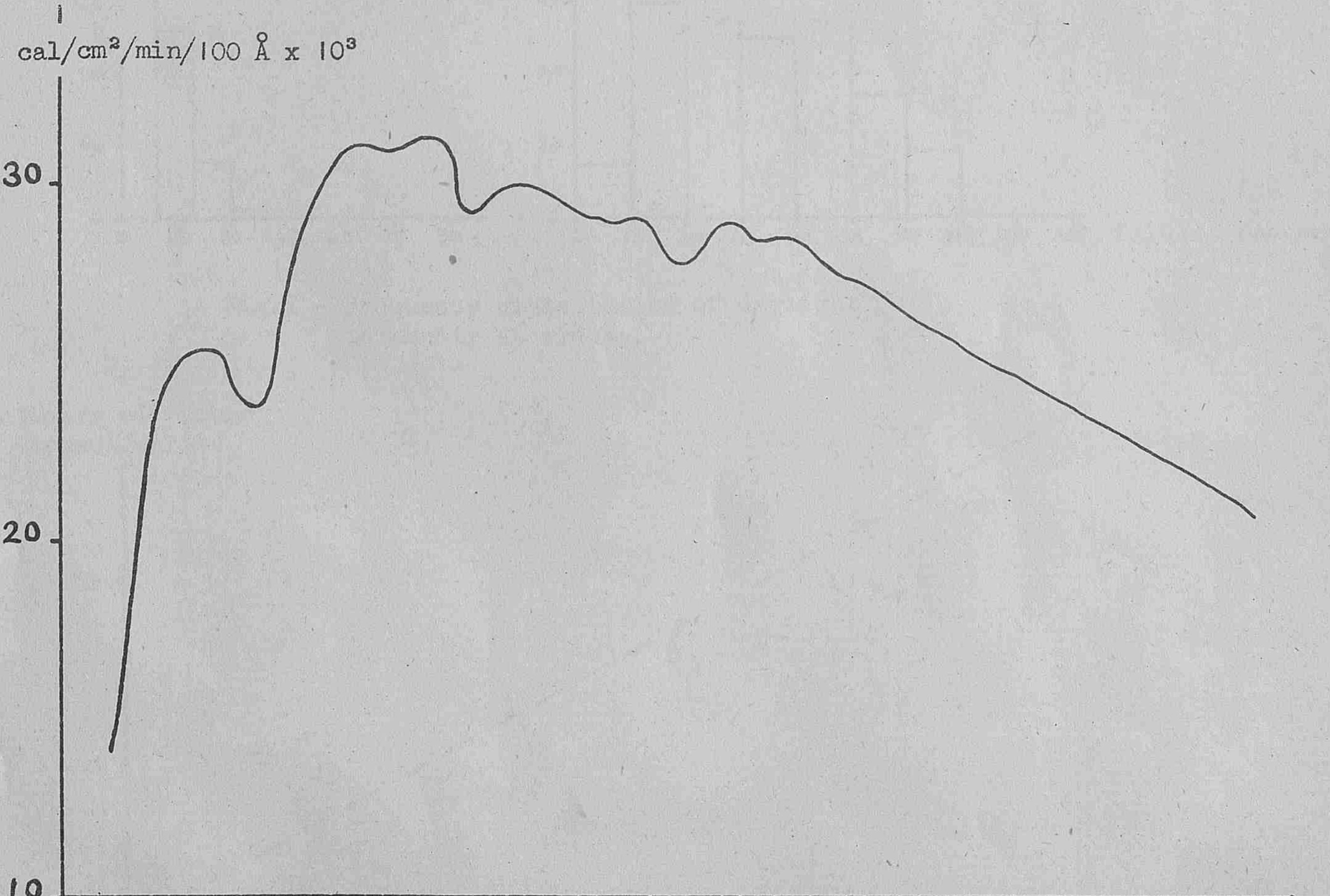


Fig.3 - Spectral distribution of extra-terrestrial solar radiation (after Nicolet).

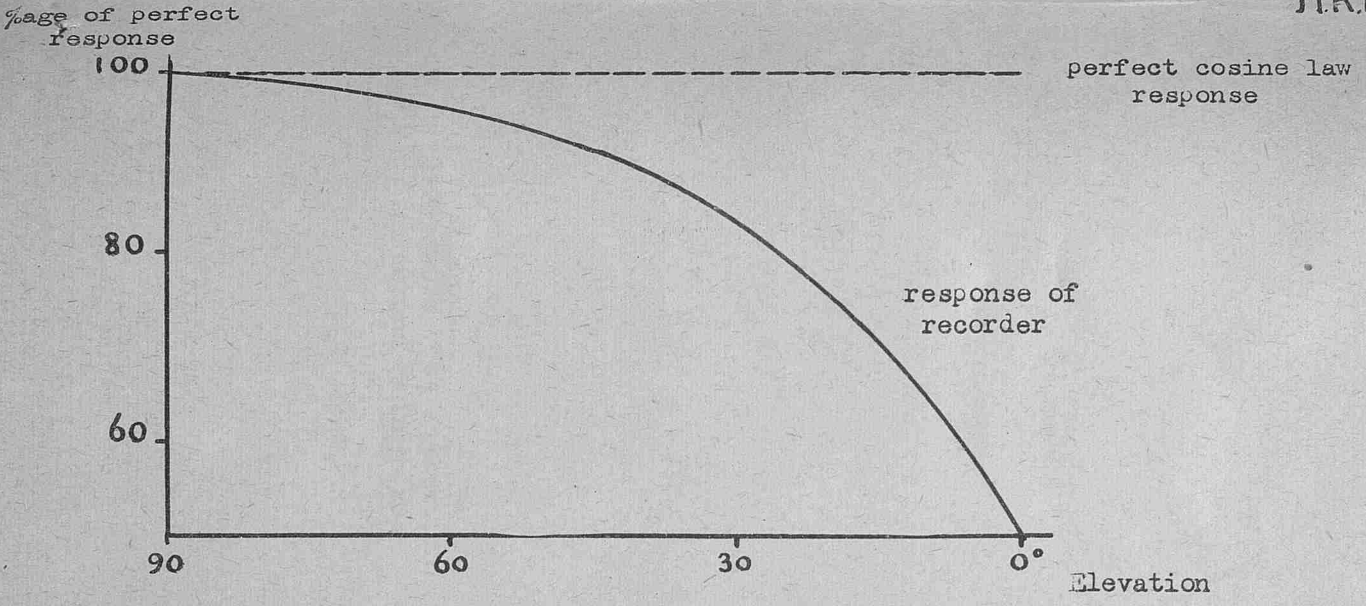


Fig.4 - Departure of response of daylight recorder from cosine law.

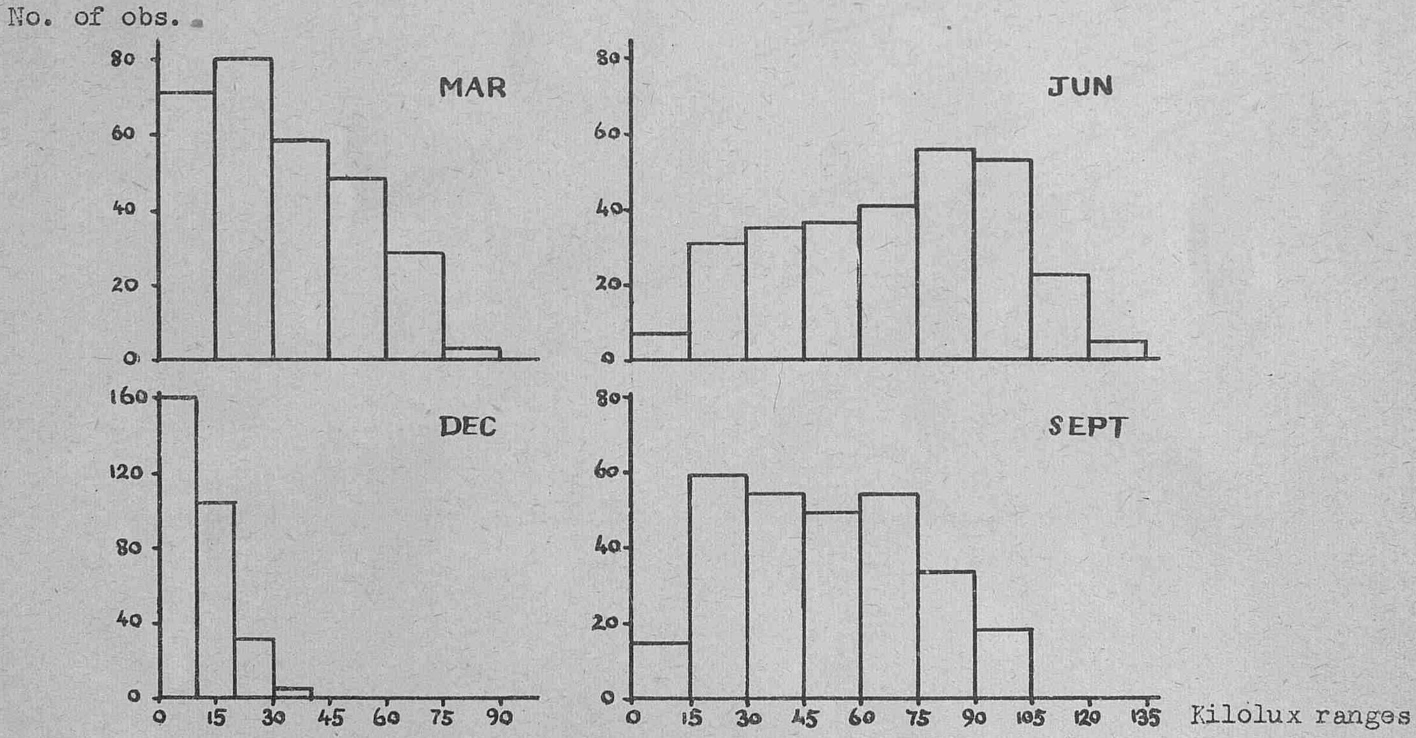


Fig.7 - Frequency distribution of daylight intensity at midday.

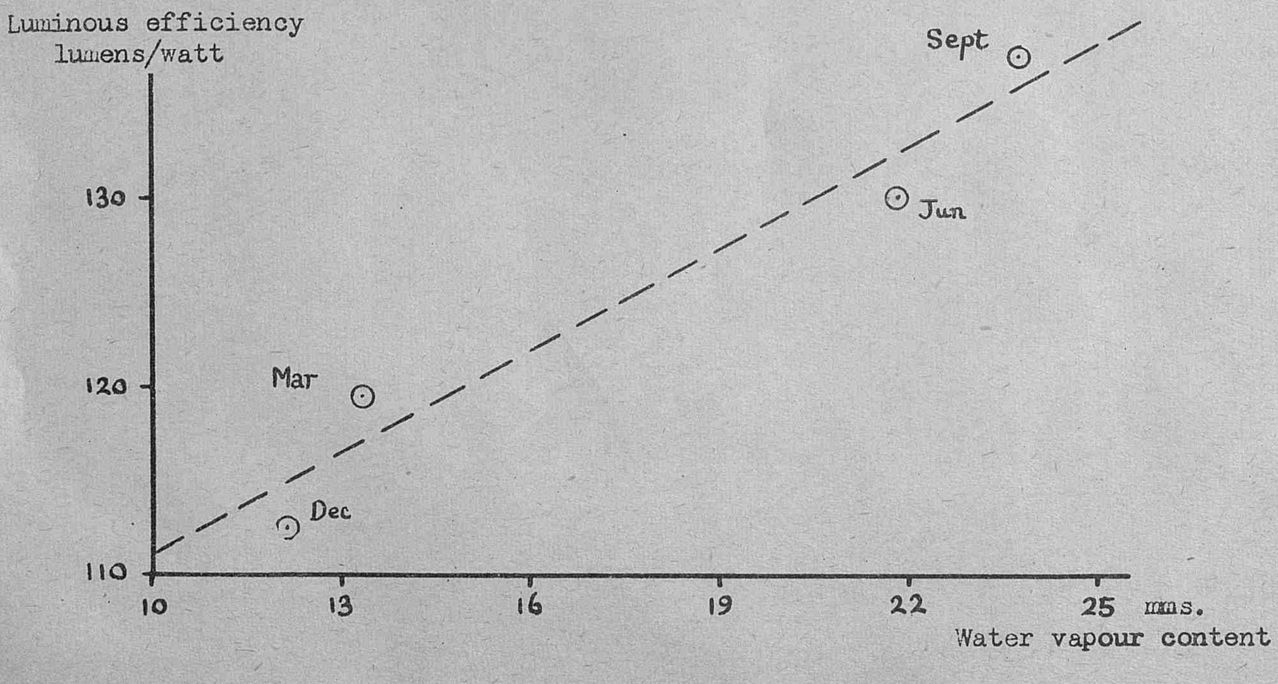


Fig.8 - Relation between luminous efficiency and water vapour content of the atmosphere.

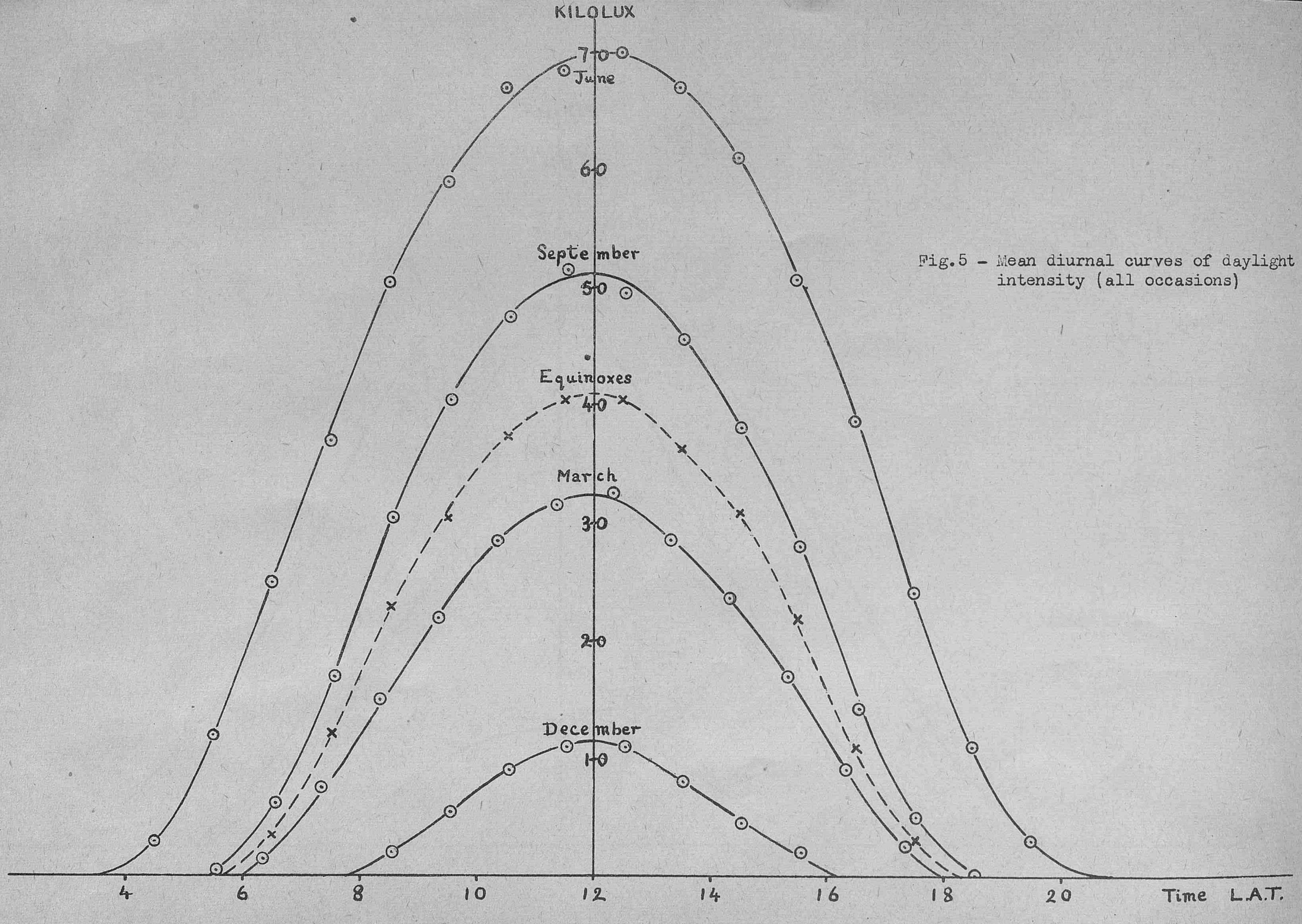


Fig.5 - Mean diurnal curves of daylight intensity (all occasions)

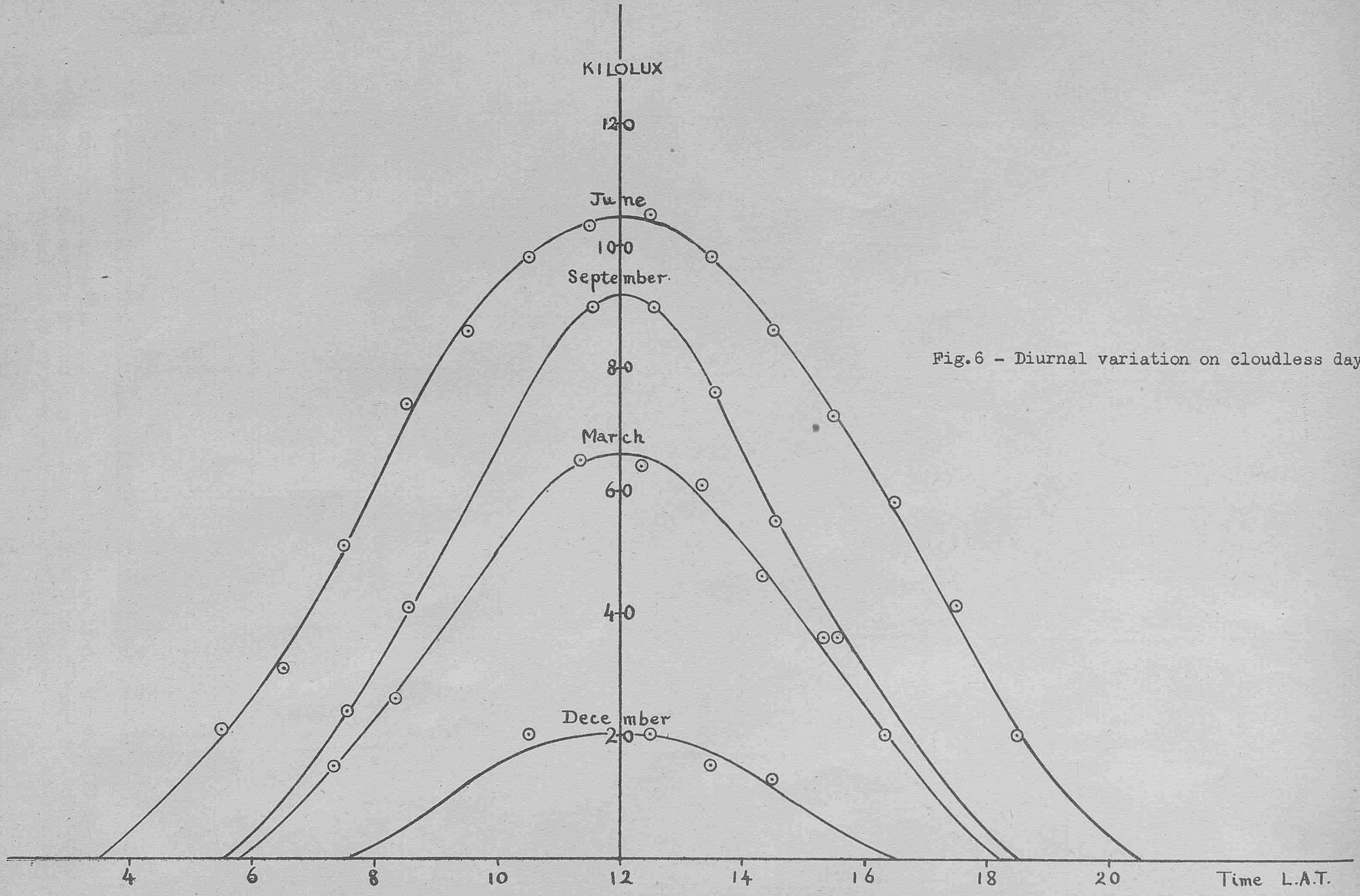


Fig.6 - Diurnal variation on cloudless days.

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M.R.P. No.833 by Mr. R.J. Murgatroyd and Mr. P. Goldsmith
A copy is available in the Library of the Meteorological Office".

Contract No. C.P. 8053.

M.R.P. 833
S.C. III/155
30th September, 1953.

AD143187

## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

# Cirrus cloud over Southern England

by
R.J. Murgatroyd
P. Goldsmith
(Meteorological Office)

## Summary

Data are presented of the frequencies of occurrence of cirrus cloud over Southern England, diurnal and seasonal variations, types of high cloud, heights and temperatures of bases and tops, thickness, visibility, and humidity in its vicinity. These data are based on surface observations and four years' high level ascents by aircraft of the Meteorological Research Flight. Details of the formation temperatures of condensation trails and the relation between their persistence and the presence of high cloud are given. Finally some mechanisms are suggested for the formation of high cloud over the British Isles.

## Introduction

Although high cloud when visible from the ground is reported in all synoptic observations and is plotted on synoptic charts, very little is known of its physical characteristics and the conditions under which it occurs. In the past, it has in fact been regarded of little practical importance compared with other types but with the advent of regular flying at cirrus levels its importance is increasing. A study has therefore been made of the observations of high cloud made during the high level ascents made by the Meteorological Research Flight from 1948-1952. These ascents were made primarily for the purpose of measuring temperature and humidity up to about 40,000 feet with cloud observations as a secondary consideration. Details of the measurements of temperature and humidity made up to 1951 are summarised in Ref.1. In 1952 particular attention was given to the high cloud observations during the flights made by one of us (P.G.) and this led to some of the conclusions given below which were also found to be consistent with the earlier observations. In addition to these flight observations use has also been made of ground observations made at the synoptic stations of South Farnborough, Shawbury, and West Raynham. A preliminary study has also been carried out on the relationship between some of the high cloud observations and the surface and upper synoptic charts for the same period.

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## Part 1. Information derived from surface observations

Surface observations of cirrus possess severe limitations. It is frequently impossible to see high cloud owing to the presence of lower clouds or fog, and observations at night are very difficult indeed. Haze layers in

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the atmosphere also have an effect. It has been noticed frequently during the flights that the amount of cirrus visible increases as the aircraft ascends and it may sometimes even happen that there is eight-eighths of thin cirrus present when none is visible from the ground. As the aircraft penetrates the thick surface haze top this increase is sudden. In addition the sun's elevation appears to be an important factor in the visual recognition of the cirrus from the surface. Nevertheless a survey of surface data on cirrus is of interest and is given below.

In Fig.1 mean frequencies throughout the day of surface observations of high cloud are shown for each month for the period 1944-52 at South Farnborough. In Fig. 2a a comparison of the daylight observations of high cloud at South Farnborough, Shawbury, and West Raynham for the years 1950 and 1951 is given. Fig. 2b is a similar diagram for daylight observations at South Farnborough only for the longer period 1944-52.

Considering Fig.1 it is seen that:-

(i) The observations all have a pronounced diurnal variation. During daylight hours cirrus is reported on an average for about 30% of the time and during the night 15% of the time.

(ii) During the winter months the cirrus observed curve rises during the morning twilight period, reaches its maximum within an hour or so, remains fairly constant throughout the day and then falls again at dusk.

(iii) During the other seasons and particularly in summer there is a peak shortly after dawn and a much larger peak around dusk. Between them there is an unsteady rise which is more pronounced in the afternoon.

(iv) The high cloud obscured curve usually has a maximum around the dawn period and a minimum around the dusk period which is less pronounced in winter. On an average the high cloud is obscured for abount 60% of the time in winter and 45% in summer.

(v) Clear skies with no cirrus were reported for about 20% of the time in daylight and 40% at night.

It is difficult to decide to what extent these observations are real. The belief that the summer dusk and dawn peaks are a visual effect is however supported by the lack of definite peaks in winter when the sun is always low. The large percentage of cases when high cloud is obscured makes it difficult to ascribe any frequency to the periods when cirrus is in fact above the station.

Considering now Figs. 2a and 2b.

(i) The curves of observations of cirrus from the surface all show maxima in the summer when values of about 40% are reached. Values of about 25% are given for all months between October and May.

(ii) The no-cirrus observations show maxima of about 25% in spring and early summer with minima of 15% in winter, with irregular fluctuations in the late summer and autumn.

(iii) The high cloud obscured curve shows a maximum of about 60% in winter and a minimum of 40% in summer. It is interesting to note that in spite of this a secondary no-cirrus minimum is found for August and September.

(iv) In Fig.2 a the curves for the three stations correspond extremely well. This would be expected as the scale of the synoptic types affecting Southern England is in general larger than the area enclosed by these three stations. This is in agreement with our aircraft observations and also from examinations of surface synoptic charts which indicate that cirrus sheets are very frequently extensive and indeed sometimes cover almost the whole of the British Isles.

(v) The overall ratios for the period of cirrus occasions to no-cirrus occasions for these stations are as follows:-

Shawbury 2.5, West Raynham 1.8, South Farnborough 1.4.

The possible explanations for this are that

(a) On the average cirrus is more likely in the west than in the east of Southern England.

(b) As Shawbury is near hills the differences may be the results of orographic lifting (Ref.2).

(c) The observations may not be representative of the actual high cloud distribution. For instance haze in the atmosphere may account for these differences.

Unfortunately there is no quantitative data available on the effective obscuration of high cloud by haze at these stations so that little can be said about (c) except that West Raynham has probably suffered most from this effect. The question of when orographic cirrus can form and the likelihood of cirrus being more frequent in western than in eastern districts of the country is discussed in Part 3 below from which it is concluded that both of these explanations could contribute in part to the high cirrus frequencies at Shawbury.

During this work a study was also made of the frequency of the various types of cirrus cloud which were reported over the same period at these stations and the results are given in Table 1.

TABLE 1. Frequency of surface reports of high cloud types

| Cirrus Code Number | Short description | % Frequency of type reported<br>South Farnboro' | Shawbury | West Raynham | Mean |
|---|---|---|---|---|---|
| 1 | Ci filaments scattered not increasing | 22 | 16 | 30 | 23 |
| 2 | Ci dense patches usually not increasing | 38 | 32 | 30 | 33 |
| 3 | Ci anvil type associated with Cu Nb | 11 | 12 | 8 | 10 |
| 4 | Ci often hook-shaped, increasing and thickening | 1 | 1 | 1 | 1 |
| 5 | Ci and Cs often in bands increasing but below 45° | 2 | 7 | 4 | 4 |
| 6 | Ci and Cs often in bands increasing above 45° | 8 | 9 | 9 | 9 |
| 7 | Cs covering whole sky | 4 | 6 | 6 | 5 |
| 8 | Cs not increasing not $^8/_8$ | 10 | 10 | 9 | 10 |
| 9 | Mainly Cc | 5 | 6 | 2 | 4 |

Types 1 and 2 are evidently reported with great frequency and Type 4 hardly at all. Types 5 and 7 are also very infrequently reported.

/Part 2.

## Part 2. Information derived from aircraft observations

(a) General description. To an observer flying in cirrus the cloud appears as a white haze. The density varies considerably and is usually greatest when the cirrus is a continuation of medium cloud. Frequently the observer is not certain whether he is in cloud or not and the isolated tufts or patches (Types 1 and 2) which are so often reported in observations from the ground seldom occur. Generally the cirrus exists in widespread sheets and these patches appear to be local increases of cloud density. The base of the cloud is usually very ill-defined and it often has streaks falling from it but the tops can be recognised more easily and it appears often that the cloud is more dense near its tops than near its base. A wave-like structure is frequently seen in the tops. Streaks have occasionally been seen coming out of the tops of the cloud but they were comparatively small and short lived.

(b) Frequency of occurrence. The aircraft observations 1949-1952 were as follows:-

TABLE 2. Total aircraft observations of high cloud

| | No. of occasions | Percentage |
|---|---|---|
| High cloud in the immediate vicinity | 112 | 46 |
| High cloud visible in the distance | 65 | 27 |
| No high cloud observed | 66 | 27 |
| Total | 243 | 100 |

There is no reason to suppose that these observations are selective in any way so that it appears that for about 50% of the time in daylight there is high cloud above any station in Southern England and that for only 25% of the time are there large areas completely clear of cirrus cloud.

If we subdivide these observations into seasons Table 3 is obtained.

TABLE 3. Seasonal observations of high cloud from aircraft

| | Percentage of occasions | | | |
|---|---|---|---|---|
| | Cirrus | Distant Cirrus | No cirrus | No. of cases |
| Winter | 58 | 10 | 32 | 41 |
| Spring | 31 | 27 | 42 | 68 |
| Summer | 50 | 33 | 17 | 70 |
| Autumn | 47 | 31 | 22 | 64 |

Unfortunately the number of winter cases is considerably less than that of the other seasons. However it appears from this table that cirrus is more frequent in summer and autumn than in winter and spring. This is in fair agreement with the data given from surface observations shown in Fig. 2a and 2b.

(c) Height of base. It has been mentioned above that the base of high cloud is very indefinite and during an ascent into it the observer notes three successive stages: (i) definitely in clear air below,

(ii) in a region of doubt "Is the aircraft in cloud or not?" (iii) now definitely in cloud. The region of doubt may in some cases extend through several thousand feet.

In Fig.3 a frequency distribution of heights of cirrus at stage (iii) is given. The values given may therefore be somewhat on the high side but are correct for all practical purposes.

It can be seen that the base of high cloud is most frequently between 25,000 and 30,000 feet and that it is usually somewhere between 20,000 and 40,000 feet in the troposphere. However there have been occasional reports of cirrus in the stratosphere. A height of 46,500 feet has been reported in Ref.3 and on a few occasions M.R.F. aircraft with a ceiling of about 40,000 feet have failed to reach the base of cirrus seen during an ascent. As these observations are of particular interest they are listed below, together with some other observations of cloud in the stratosphere made by other aircraft based at South Farnborough.

| Date | Time | Aircraft | Height of Tropopause (feet) | Maximum height of Aircraft (feet) | Estimated height of cloud base (feet) | Description of cloud |
|---|---|---|---|---|---|---|
| 4.5.53 | 1500Z | Mosquito | 39,000 | 35,000 | - | Some Cs well above 35,000' may be in the stratosphere. |
| 5.5.53 | 1000Z | Mosquito | 38,000 | 41,000 | 45,000 | Thick dense extensive Cs. |
| 1.7.53 | 1600Z | Valiant )<br>Canberra)<br>Meteor ) | 38,000 | - | - | Cumulonimbus top at 42,000'. |
| 27.7.53 | 1400Z | Meteor | 28,000 | 40,000 | - | Some Ci above 40,000' at about 30 miles west of Farnborough. |
| 29.7.53 | 1200Z | Canberra | 34,000 | 42,000 | - | A small amount of Ci or contrail above to the north of Farnborough. |
| 30.7.53 | 1500Z | Mosquito | 35,000 | 38,000 | 45,000 | A layer of thin Cs to the SW of Farnborough. |
| 7.8.53 | 1500Z | Mosquito | 36,000 | 39,000 | 45,000 | Thin Cs thickening. |

(d) Height of tops. On most of the occasions examined the high cloud appeared to extend to the region of the tropopause. In view of the variability of tropopause height it was therefore attempted to relate high cloud tops to tropopause level. This has been done in Fig.4.

It appears that the tops of the cirrus cloud are most frequently less than about 2,000 feet from the tropopause. It is not known to what extent this conclusion is determined by the nature of the definition of the tropopause, but in practice it appears that the first definite decrease in the lapse rate in the tropopause region determines the tops of the cloud. A few cases have been reported when the tops apparently extend into the lowest 2,000 feet of the stratosphere. On other occasions the tops may be as much as 11,000 feet below the tropopause. In these cases the cloud is often patchy, in contrast to the usual widespread effect and resembles patchy altostratus in appearance and density.

(e) Cloud thickness. A frequency distribution of cloud thicknesses is given in Fig.5. Cirrus cloud most frequently has a thickness between 3,000 and 5,000 feet but thicknesses up to 12,000 feet have been observed. Sometimes it merges with the medium cloud. When the thickness is great there is a tendency to layering, but it is difficult to decide whether in fact clear lanes exist with cirrus above and below or alternatively there merely is a stratification of cloud density.

(f) Relationship between the height of the defined tropopause and the base and thickness of cirrus cloud. Fig.6 shows a plot of all the observations of base and top height plotted against the height of the tropopause. Although the number of observations for each of these heights is small and their scatter is large an attempt has been made to draw lines of mean base and top levels for each tropopause height. The following tentative conclusions are drawn from these lines:-

(i) The mean tops appear to reach the tropopause for heights up to 30,000 feet but at higher levels they progressively become relatively lower until at 40,000 feet they are on average about 4,000 feet below the tropopause. It is considered that this happens because as the tropopause height increases, so does the chance of an abrupt slackening of the lapse rate in the high troposphere become greater resulting in the limitation of the cirrus tops.

(ii) The mean bases tend to be at a constant thickness below the tropopause up to a tropopause height of 30,000 feet. At higher levels however the mean base appears always to be in the region of 30,000 feet. Consequently on the average, mean cloud thickness with tropopause levels below 30,000 feet is about 4,000 feet and it progressively increases to 7,000 feet with a tropopause level of 40,000 feet.

We would emphasize that the above conclusions refer only to mean values. Individual cases as can be seen from the scatter of the points on the diagrams may differ widely from them.

(g) Temperature of base and tops. A frequency distribution of temperature of the base of high cloud is given in Fig. 7. The most frequent values are from -40 to -50°F. but measurements have been made within the range -10 to -80°F.

The corresponding frequency distribution of cloud top temperatures is given in Fig.8. This distribution is rather flat between temperatures of -40 and -75°F. but a few values have been measured between -20 and -30°F. and down to -100°F. The lack of temperatures between -30 and -40°F. is interesting and suggests that the few values between -20 and -30°F. belong to another distribution - the low cirrus which is very similar to medium cloud.

The temperature of -40°F. is about the mode of the base observations and there are very few observations of high cloud tops at temperatures greater than -40°F. This is also the temperature given by many laboratory investigators at which spontaneous nucleation becomes effective. If there is a paucity of ice-forming nuclei at high levels as appears likely (Ref.4) it could easily happen that the majority of cirrus forms at -40°F. or below without the assistance of these nuclei and that subsequent or self nucleation could lower the bases to include higher temperatures.

The possibility of formation of cirrus cloud however includes considerations of the moisture available as well as the likelihood of nuclei being present and this point is considered in a subsequent paragraph.

(h) Humidity in and around high cloud. A frequency distribution of depression of frost point relative to air temperature as measured with the visual pressurised Dobson-Brewer hygrometer during the periods when high clouds were reported is given in Fig.9. The values used in each case were the minimum depression of frost point within or near the cirrus layer, the readings being taken at 1,000 feet intervals. It has previously been thought that at very low values of frost points the uncertainty of measurement was only about 4°F., and the accuracy somewhat better in cirrus cloud. Recent flights with automatic hygrometers at cirrus levels however have shown that short period fluctuations in the frost point of 5 to 10°F. within distances of less than a mile are common. It is impossible to measure these fluctuations with the visual type of instrument which only gives a mean value over a distance of several miles.

The air temperature can probably be measured to about 1°F. at these levels. Hence it appears from the measurements with the visual instrument that in the immediate neighbourhood and even within cirrus cloud the air is not always saturated with respect to ice but that on an average there is a frost point depression of about 3°F. (about 80% relative humidity), and that lower values of relative humidity are common. However, according to Ref.4, saturation with respect to water i.e. considerable supersaturation with respect to ice is necessary for cirrus cloud formation.

It has also been noted that it is possible to form condensation trails in an atmosphere 60% to 80% saturated and unless cirrus cloud is present these trails almost invariably disappear within minutes. Furthermore it appears likely on theoretical grounds that ice crystals of the size measured in cirrus cloud would evaporate within times of this order in an atmosphere only partially saturated. Hence there is a difficulty in explaining why cirrus cloud should persist at all if these conditions were constant. The fluctuations revealed by the automatic hygrometer may be sufficient to provide regions of supersaturation. Further data is required before these comparatively low mean values of humidity can be reconciled to the long period of persistence of the cirrus cloud sheets as a whole.

(j) Humidity distribution when high cloud is present - the "Moist Layer". When high cloud is present or upwind in direction the distribution of frost points with height usually exhibits the characteristic form shown in Fig. 10a. This example was selected from the "cirrus in the distance" cases, and shows a rapid increase of humidity at 28,000 feet with a moist layer persisting to 32,000 feet, the height of the tropopause. This "moist layer" becomes more pronounced as the distance from the cirrus is decreased until, when cirrus is present near the aircraft, the sounding takes the form of the example in Fig. 10b. The moist layer is not always easy to trace and its depression relative to the dry bulb temperature may be as much as 15-20°F., but this depression quickly decreases as the cloud approaches. When no cirrus is observed the sounding frequently takes the form shown in Fig. 10c. A network of ascents giving this type of information would thus be invaluable to a forecaster who is required to give detailed information on high cloud.

It is necessary at this stage to qualify the above remarks on account of the different ways in which cirrus cloud may be formed. This is discussed further in a subsequent section. It appears that most cirrus occasions arise when a warm ridge of air recently over the Atlantic is advected over the British Isles at cirrus levels. Other cirrus occasions however arise associated with cold pools or strong convection and the moist layer is not then evident.

Regarding the moist layer as potential cirrus cloud and plotting all the bases and tops relative to the tropopause Fig.11 was obtained. The tops of the moist layer tend to coincide with the tropopause for low tropopause heights but become somewhat lower as the latter increases. Mean moist layer bases tend to have the same distribution with height as the cirrus bases shown in Fig.6 and never exceed 30,000 feet. Consequently mean thickness also increases with height.

?? As the mean base is about 30,000 feet and its extreme value rarely below 24,000 feet its temperature is generally between -25° and -45°F. Similarly the tops are rarely below 30,000 feet - i.e. at a temperature of -45°F. Hence, as has already been stated in paragraph (g) above, high cloud is potentially always able to form at temperatures below -40°F. and therefore it does not seem necessary to postulate the presence of ice nuclei at cirrus levels in order for the cloud to form.

Cirrus cloud is much more frequent than medium cloud. It appears that this can arise both from the fact that the necessary moisture is more often available at the higher levels and also from the fact that at medium cloud levels temperatures are higher and therefore require some form of nuclei to assist in the condensation process.

(k) Visibility through high cloud. No observations of visibility in cirrus are recorded, but the following details are based on the experience of flying in high cloud.

High cloud usually limits the visibility from air to ground but only when the high cloud is a continuation of medium cloud is it likely that the surface will be completely obscured. Normally the ground (or low cloud) is still clearly visible for several miles ahead when the aircraft is flying in high cloud, indeed sometimes the cloud is so thin that its presence makes little or no difference to vertical visibility. However, vertical visibility is often more limited when the aircraft is above the high cloud. This effect is similar to that of surface haze; when flying just above surface haze the ground is often more indistinct than when actually in the haze layer.

The horizontal visibility in high cloud is a more difficult factor to estimate, but when aircraft of the M.R.F. have been formating in high cloud, although the horizontal visibility was more limited than in clear air, this limitation was not serious and no difficulty was experienced.

## Part 3. Notes on the forecasting of (a) high cloud and (b) condensation trails

### (a) High cloud.

(i) At present a forecaster has not many tools which help him to forecast the probable occurrence of cirrus cloud. The classic model of fronts, particularly warm fronts, provides one type of phenomenon which may be associated with high cloud. However it is apparent from daily examination of synoptic charts that the presence of the front is neither a necessary nor a sufficient criterion to use. High cloud frequently occurs without being associated with fronts and on the other hand surface fronts need not extend to cirrus levels.

It is the purpose of this section to suggest other factors which can usefully be taken into consideration when it is necessary to provide a forecast of high cloud.

(ii) The general character of the weather at any point in the lower troposphere is determined by the nature of the air masses involved and the fronts which occur at their boundaries. It is not easy to extend the idea of air mass to cirrus levels but it is usually possible to gain an idea of the recent history of the air there by studying the 300 mb. contour charts. Thus to some extent we can say that the air over the British Isles at 300 mb. is of recent polar or Atlantic track, although of course borderline cases are frequent. It would perhaps be more profitable theoretically to discuss the air characteristics at this level in terms of the position of the long wave pattern with relation to the area involved but this may not be easy in practice as shorter wave phenomena no doubt would be important as regards cloud formation.

It is considered that air at 300 mb. which has had a recent polar history is in general more likely to be free of cirrus than air of a recent Atlantic history. This conclusion is supported by (1) a short study of high cloud frequency with respect to wind direction at cirrus levels and (2) an examination of the moisture content at these levels with respect to the recent history of the air. Describing these now in turn:-

1. From considerations of the relationship between cirrus occurrence and the 300 mb. wind direction.

Over the two years 1950-1 all surface observations from South Farnborough of sky obscured, no high cloud, and high cloud present have been counted for winds in the eight sectors of the compass. The wind directions used were the generally prevailing directions given by the Larkhill ascents between about 20,000 feet and the tropopause. As the ascents were six-hourly and the cloud observations hourly the wind direction was assumed to remain in a given octant for the three hours preceding and following an ascent.

The results are given in Table 4 which shows the percentage of occasions for each octant in which the sky was obscured, there was no high cloud and high cloud was observed.

TABLE 4. The relationships of high cloud observations to prevailing wind direction in the upper troposphere

| Wind directions | 000° - 040° | 050° - 090° | 100° - 130° | 140° - 180° | 190° - 220° | 230° - 270° | 280° - 310° | 320° - 350° |
|---|---|---|---|---|---|---|---|---|
| | Percentage of occasions | | | | | | | |
| Sky obscured | 51 | 64 | 59 | 62 | 55 | 52 | 49 | 52 |
| No high cloud | 39 | 32 | 34 | 22 | 20 | 23 | 27 | 32 |
| High cloud | 10 | 4 | 7 | 16 | 25 | 25 | 24 | 16 |
| Ratio $\frac{\text{high cloud}}{\text{no high cloud}}$ | .3 | .1 | .2 | .7 | 1.3 | 1.1 | .9 | .5 |

From this table it appears that the percentage of high cloud occasions is greatest when the wind is in the SW to W sectors and least in the NE to E sectors. The percentages of no high cloud exhibit the reverse trend.

Considering the ratio of high cloud to no high cloud it is seen that, irrespective of the fact that westerly winds at 300 mb. are much more frequent than easterly winds at that level, the westerly winds are more likely to produce cirrus. It may be noted also that as easterly winds which are mainly continental in track are unlikely to produce cirrus it is very improbable that nuclei of continental origin are of any importance as regards the production of cirrus cloud over Southern England.

2. From considerations of the recent history of the air at cirrus levels and moisture available.

From the high level ascents made by the Meteorological Research Flight and the relevant 300 mb. contour charts it appears that air at cirrus levels which can be associated with contours extending over large areas of the Atlantic usually has the requisite moisture content. In fact out of the 126 "moist layer" cases found, 106 were in air with a recent Atlantic track, 12 with a recent track over polar regions and 2 with a continental track. There were 9 Atlantic, 12 polar and 2 continental cases with no moist layer. In addition there were about 30 cases with very low tropopauses or cold pools whilst a number of ascents had to be rejected from this study for lack of data. Air streams of recent polar and continental history frequently do not have sufficient moisture at these levels. Hence on this account most cirrus is likely to be associated with air of recent Atlantic history.

(iii) Although it is thought that the recent history of the air exerts a considerable influence on its potentiality to form high cloud other factors which may tend to produce the necessary ascent to bring about condensation must receive full consideration when producing a forecast. A complete list of these factors has not been produced but those which appeared most important from a short study of synoptic cases for the British Isles in 1952 are as follows:-

1. The presence of a jet stream over the area.
2. Warm advection at cirrus levels.
3. Cold pools at cirrus levels.
4. Cirrus formation from tops of cumulonimbus clouds.
5. Orographic lifting.

No doubt this list could be extended and the relative importance of the various factors assessed by means of a comprehensive synoptic study, but this is beyond the scope of the present paper.

Short notes are given below about each of these factors in turn.

(iv) It has been noticed that the axis of a jet stream frequently coincides roughly with the edge of the cirrus sheet. For this purpose the jet stream speed need not be in the order of 100 knots but merely a wind speed maximum of 50 knots or so. On the cold side there is usually very little high cloud whilst on the warm side cirrus is often extensive. This has also been described by Murray (Ref.5) in a report giving an analysis of jet stream flights made by M.R.F. and by Sawyer and Betty Ilett (Ref.6) in a paper describing a statistical examination of cloud distribution near the jet stream. Schaefer (Ref.9) has also associated the occurrence of high clouds with the jet stream in the U.S.A. In our experience the rough boundary of the cloud and no-cloud areas at the jet stream axis is most well marked with jet streams containing a considerable northerly component, cirrus then often being present over western and not over eastern districts of the British Isles.

(v) The classical conception of warm fronts has always included cirrus cloud well ahead of the surface front and in fact it has been found that it can often be associated with surface fronts up to 1,000 miles away. Between the 500 and 300 mb. levels this situation usually shows advection of a warm ridge. The nature of this mechanism by which the necessary cooling for condensation occurs with this warm advection is not fully understood. It was formerly thought to be due to general up-slope motion on the frontal surface but this picture is complicated by the presence of the jet stream which is so often present in the upper troposphere near the junction of air masses, and it is now thought by some authorities that frontal surfaces in fact do not usually extend up to cirrus levels. If the cirrus formation is related to warm advection in this way when warm surface fronts are present there appears to be no prima facie reason why it should not occur when they are absent provided that the situation in the 500-300 mb. layer is broadly similar. A short study of synoptic cases suggested that this is almost always so and warm front cirrus formation may be regarded as a particular case of cloud produced by warm advection in the upper troposphere. In the cases where the connection was not obvious it appeared that the cloud was usually ahead of the main warm advection area. Cold advection and areas of no high cloud appear to be similarly related.

There is considerable support for this idea in general in Ref.(8). The results of M.I.T. Pressure Change Project showed that there was a strong correlation between the sign of the vertical motion and the sign of the temperature advection at 700 mb. and above, warm air advection being positively correlated with ascending air and conversely.

It is also shown in Ref.(7) that changes in the thickness of the 1,000-700 mb. and the 700-500 mb. layers due to advection are largely offset by those due to dynamic heating or cooling or non-adiabatic processes. This compensation was found to be more complete in the higher layer. If this conclusion can be extended to the 500-300 mb. layer it appears quite feasible that warm advection there can be related to a process resulting in dynamic or other cooling and hence a tendency for high cloud to form.

(vi) It was also noted that cold pools and deep troughs on the 500-300 mb. chart are often accompanied by large areas of high cloud. In the case of the troughs the high cloud is generally to the east and south-east and is often associated with warm advection round the south of the trough. With cold pools the high cloud is often more widespread.

(vii) Cirrus cloud may of course often be formed from the tops of cumulonimbus clouds. In such a situation it is often patchy and localised. The cirrus found in conjunction with cold fronts is frequently of this type and is usually not extensive.

(viii) Orographic lifting has also been suggested as a mechanism for forming cirrus cloud (Ref.2). It can be demonstrated however that the conditions when this can occur are somewhat limited. From Ref.(1) it appears that on the average, i.e. including cirrus and non-cirrus cases, the depression of frost point with respect to temperature at cirrus levels is about 10 to 12°F. Our Fig.9 shows a mode of 2 or 3°F. depression in and around cirrus cloud. Therefore a temperature loss of 8° or 9°F. due to orographic lifting to reach saturation with respect to ice is usually necessary. For saturation with respect to liquid water the cooling would have to be about another 6°F. i.e. about 15°F. in all. With adiabatic conditions of lifting, therefore, an upward displacement of about 3,000 feet would be required on the average and often considerably more to produce cirrus by the orographic mechanism. In general a lift of 1,000 to 2,000 feet would only be exceeded infrequently and the air at cirrus levels would have to be almost saturated with respect to ice for it to be effective in producing cloud. It is thought therefore that this mechanism would be effective only on the occasions when there is a "moist layer" i.e. usually in air of recent Atlantic track, and either on the warm side of a jet stream, in the vicinity of a cold pool or ahead of a warm front or other warm advection. The effect of orographic lifting would then be most evident in cases where high cloud would be expected from these other considerations and would usually either advance its onset or produce local thickening in the cirrus layer. All the cases discussed in Ref.(2) can be explained on these grounds.

(ix) Although the above factors are helpful in assessing the likelihood of high cloud occurrence there are many synoptic situations when they cannot be used. Cases with no jet stream or cold pool and thickness lines parallel with the wind directions or contours are difficult to assess as also are those associated with small thickness gradients or with concentric contours and thickness lines. The cirrus cloud in inverted situations with contour or thickness highs to the north of the British Isles are often difficult to explain by the above factors. A further study preferably on a statistical basis of the occurrence of high cloud with relation to these and other synoptic factors will be necessary before high-cloud forecasting can be achieved with confidence in these latter situations.

(b) Condensation trails. Condensation trails are usually formed at high cloud levels and they may be regarded as a form of high cloud. Their process of formation is of course an artificial one but once they are formed their behaviour is governed by the same processes as for natural cloud. Thus if atmospheric conditions are such that cirrus cloud is present we might expect condensation trails at the same level to be persistent. Furthermore if the moisture content of the air is large the trails are likely to be longer and of greater duration than if it is small.

These considerations have been examined with respect to the contrails and clouds observed on the high level ascents and the results are presented in Fig.12. Frequency distributions of contrail persistence on occasions of cirrus, distant cirrus, and no cirrus are given. The diagrams illustrate the following points:-

1. When cirrus cloud is present very persistent contrails are the rule. The frequency of 65% given for Contrails Type 7 (very persistent) is almost certainly low. The reason for this is that on occasions it may happen that cirrus is present at lower levels than those at which it is possible for the condensation trails to form from considerations of the relative positions of the temperature sounding and the "Immunity Line" on the Te-Phi-Gram.

2. When distant cirrus or no cirrus is reported very persistent contrails are rare (less than 10%).

/3.

3. The mode for the no cirrus diagram is Type 3 Contrails and that for distant cirrus Type 4. - the latter being the longer. It has been shown above that distant cirrus is usually associated with a moist layer and therefore it appears that the length of a non-persistent trail is a function of the humidity.

Our experience up to the present has provided the following information on which forecast of contrails can be based:-

(a) Contrails normally form for piston-engined aircraft (e.g. a Mosquito) at an ambient temperature of about 12°F. below the Immunity Line (Mintra) as printed on the present Meteorological Office's Te-Phi-Gram (Form 2810A), irrespective of humidity. For jet aircraft it appears from the relatively few experiments that we have yet made that the corresponding figure should be -19°F. i.e. jet aircraft will normally start to form contrails at about 2,000 feet (depending on the lapse rate) higher than a piston engined aircraft. These figures are only averages and may be varied a few degrees by the engine setting of the aircraft. Under suitable marginal conditions contrails can be switched on or off by varying throttle settings.

(b) If cirrus is present and the difference between the Mintra line and the ambient temperature is sufficient for the contrails to form then they will be very persistent at the high cloud level. At other levels they will be non-persistent but if the cirrus cloud coincides with the start of the contrails, then they will be persistent immediately without having a layer of non-persistence below.

(c) Contrails normally appear to start becoming shorter just above the tropopause and short trails have still been observed several thousand feet into the stratosphere. More precise information on the conditions at which contrails finally stop in the stratosphere has not yet been obtained owing to the limited ceiling of the aircraft so far available.

As very persistent contrails are more easily visible from the ground than thin cirrus, then, when the former are observed, it can be assumed that the latter is present in extensive sheets even though it cannot be seen from the ground.

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(4) 1948. Ludlam, F.H. Quart. J.R.Met. Soc., Vol. 74.

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(9) 1953. Schaefer, V.J. Tellus, Vol. 5, No.1.

M.R.P. 833.

**MEMO ROUTING SLIP**

NEVER USE FOR APPROVALS, DISAPPROVALS, CONCURRENCES, OR SIMILAR ACTIONS

| | | |
|---|---|---|
| 1 NAME OR TITLE | INITIALS | CIRCULATE |
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| 4 | | SEE ME |
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REMARKS

MRP 933
MRP 833 (or Prof Note 118)
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DD FORM 1 FEB 50 95 Replaces DA AGO Form 895, 1 Apr 48, and AFHQ Form 12, 10 Nov 47, which may be used. 16—48487-4 GPO ☆

Graphs showing the Mean Monthly Frequencies of Surface Observations of High Cloud throughout the Day over the Pe

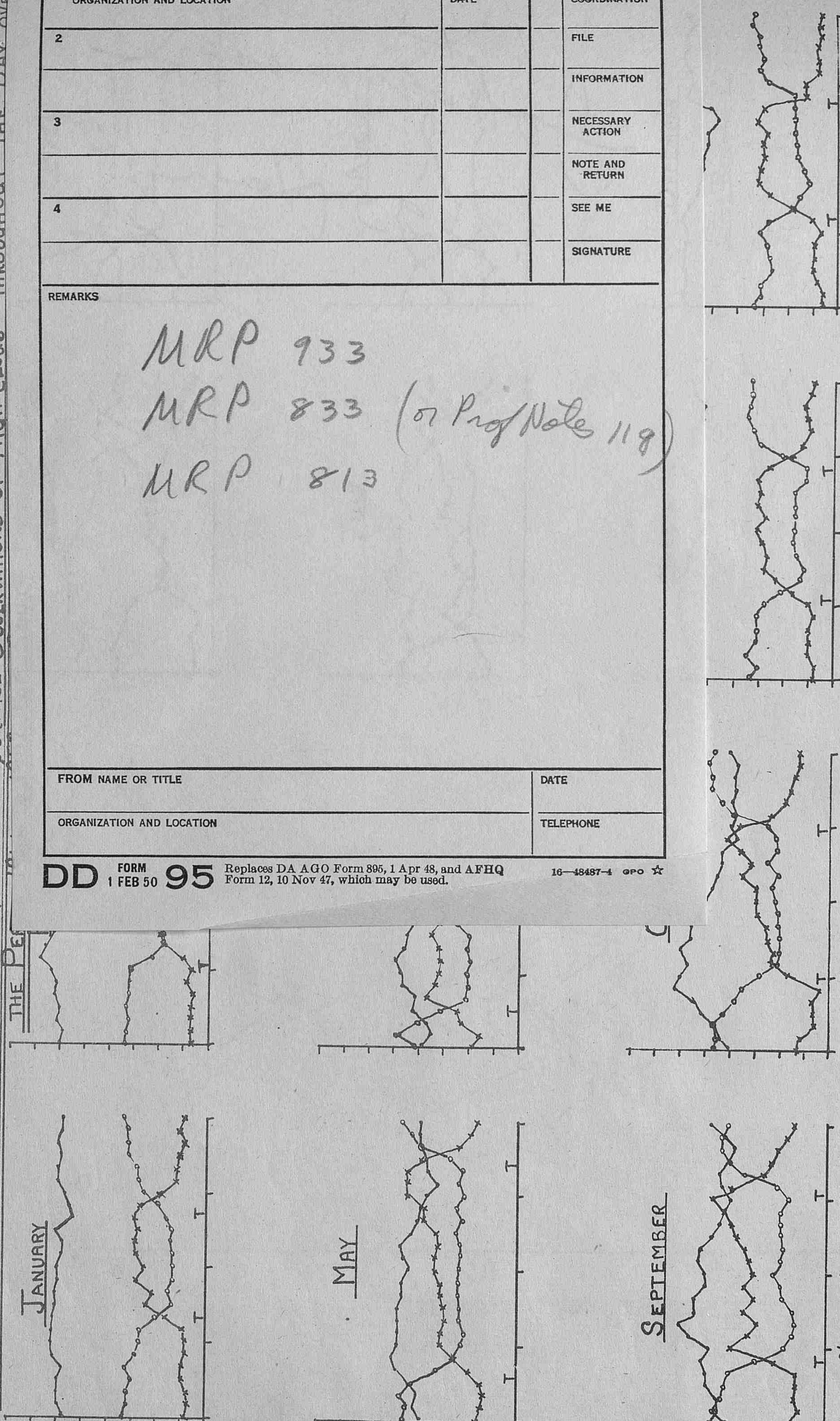


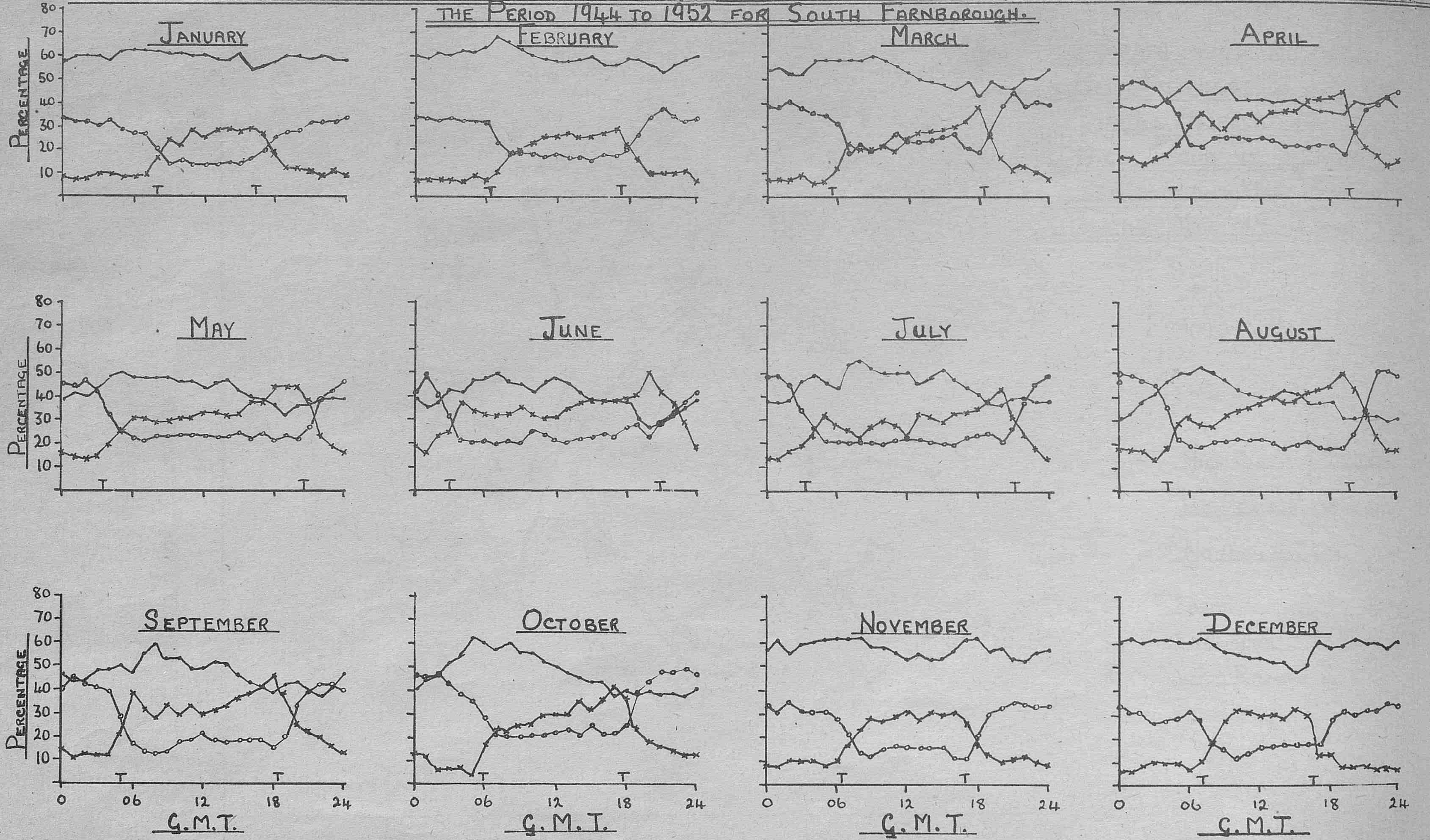
Figure 1. —•— High Cloud Obscured, —o—o— No High Cloud, —x—x— High Cloud Observed.

T. Denotes the approximate times of twilight.

Johannsen

25 Jan 57

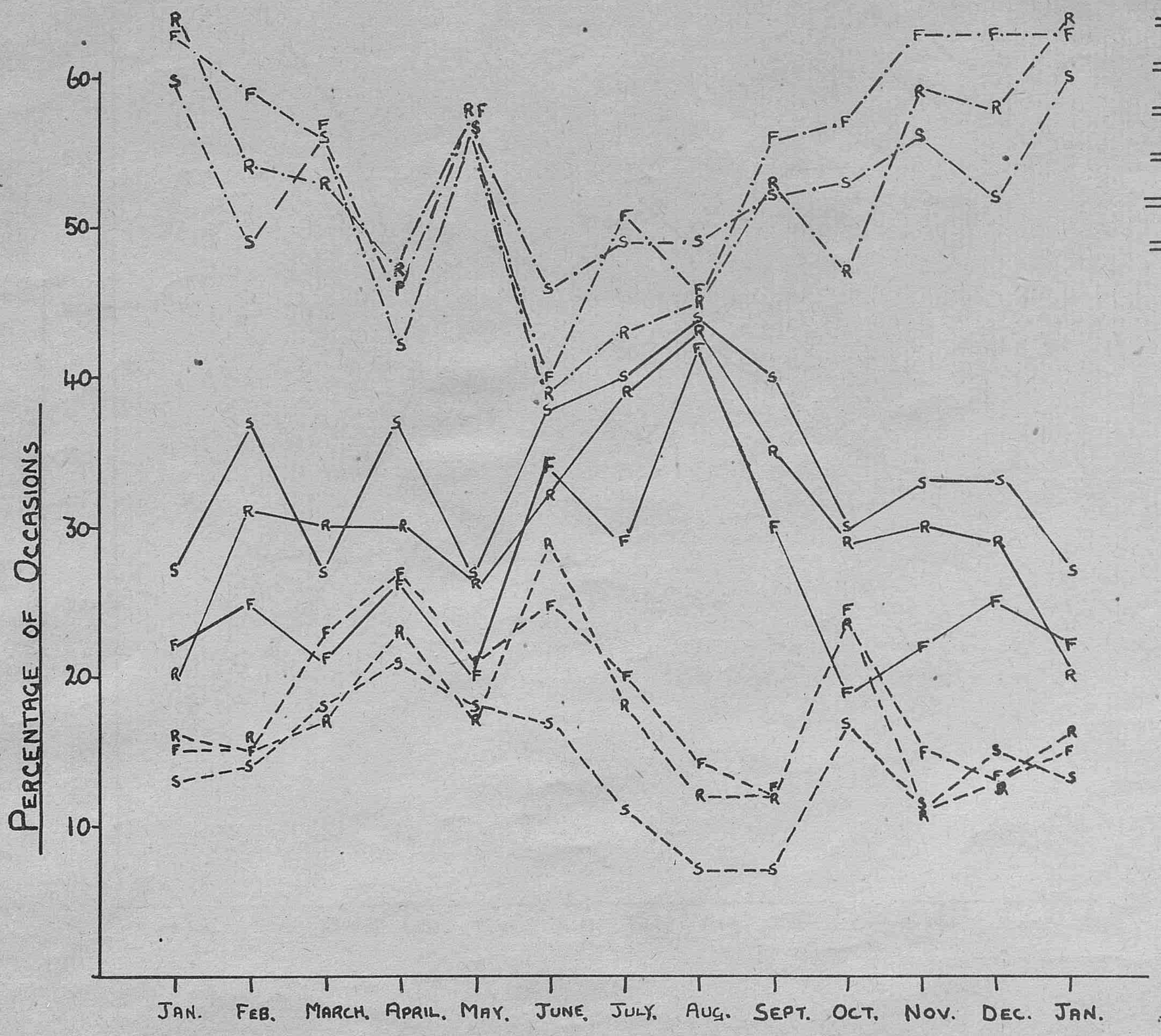
# GRAPHS SHOWING THE MEAN MONTHLY FREQUENCIES OF SURFACE OBSERVATIONS OF HIGH CLOUD THROUGHOUT THE DAY OVER THE PERIOD 1944 TO 1952 FOR SOUTH FARNBOROUGH.



T. DENOTES THE APPROXIMATE TIMES OF TWILIGHT.

FIGURE 1. •—• HIGH CLOUD OBSCURED, o—o—o NO HIGH CLOUD, x—x—x HIGH CLOUD OBSERVED.

# A Graph showing the Mean Frequency of Day-time Surface Observations of High Cloud at S. Farnborough, Shawbury and West Raynham for the Years 1950 and 1951.



-·-·-·- High Cloud Obscured.
——— High Cloud Observed.
- - - - No High Cloud.
F Farnborough.
S Shawbury.
R West Raynham.

Figure 2. a.

# A GRAPH SHOWING THE MEAN FREQUENCY OF DAY-TIME SURFACE OBSERVATIONS OF HIGH CLOUD AT S. FARNBOROUGH FOR THE YEARS 1944 TO 1952.

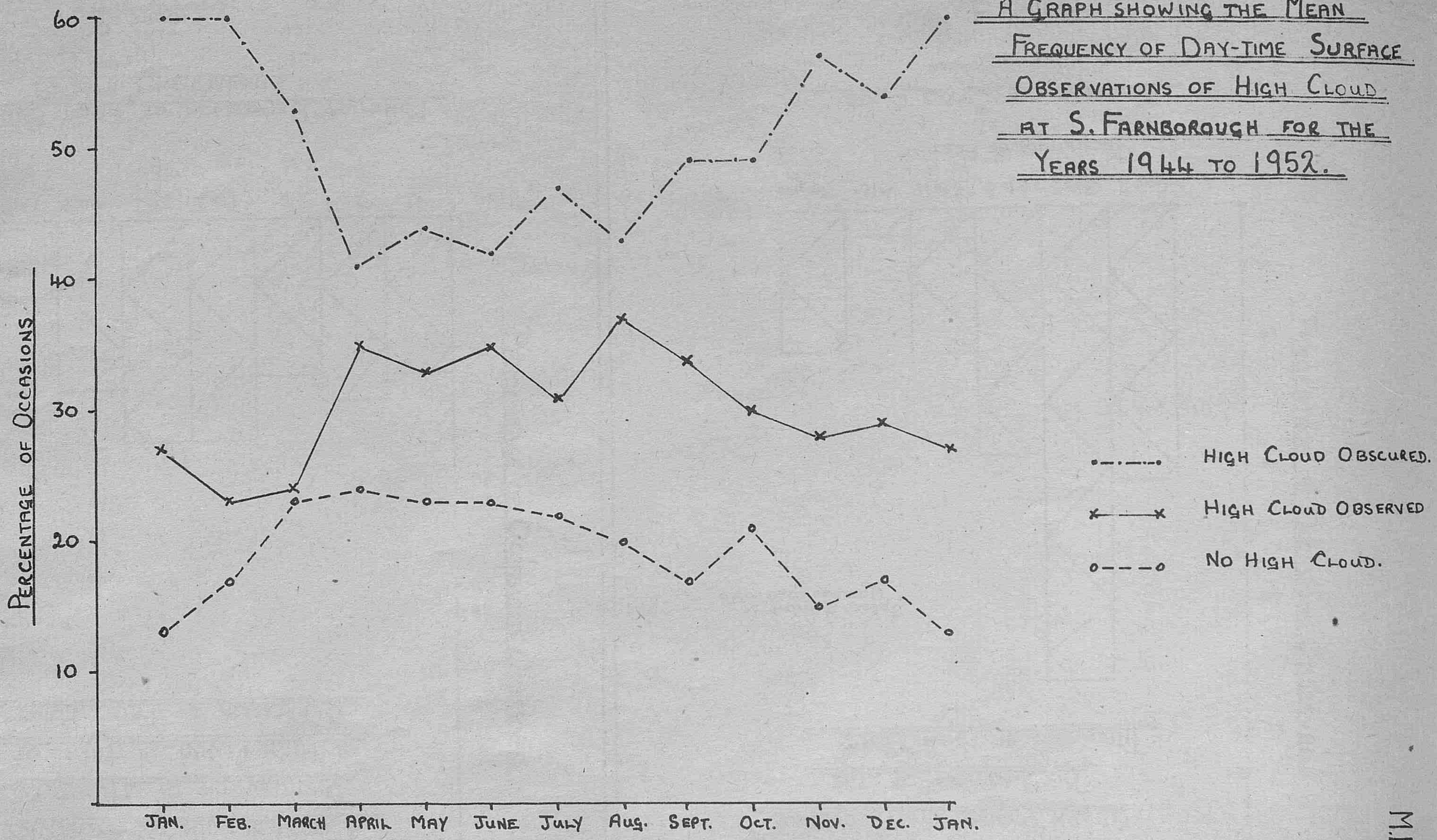
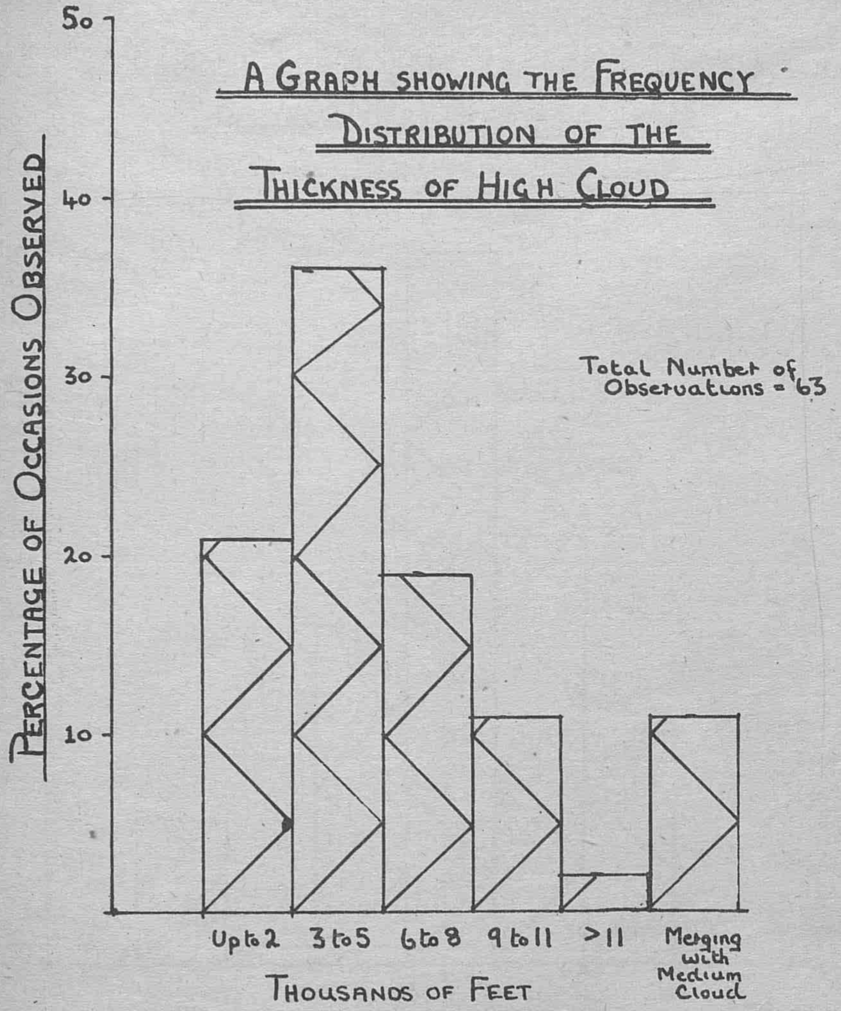


FIGURE 2.b.

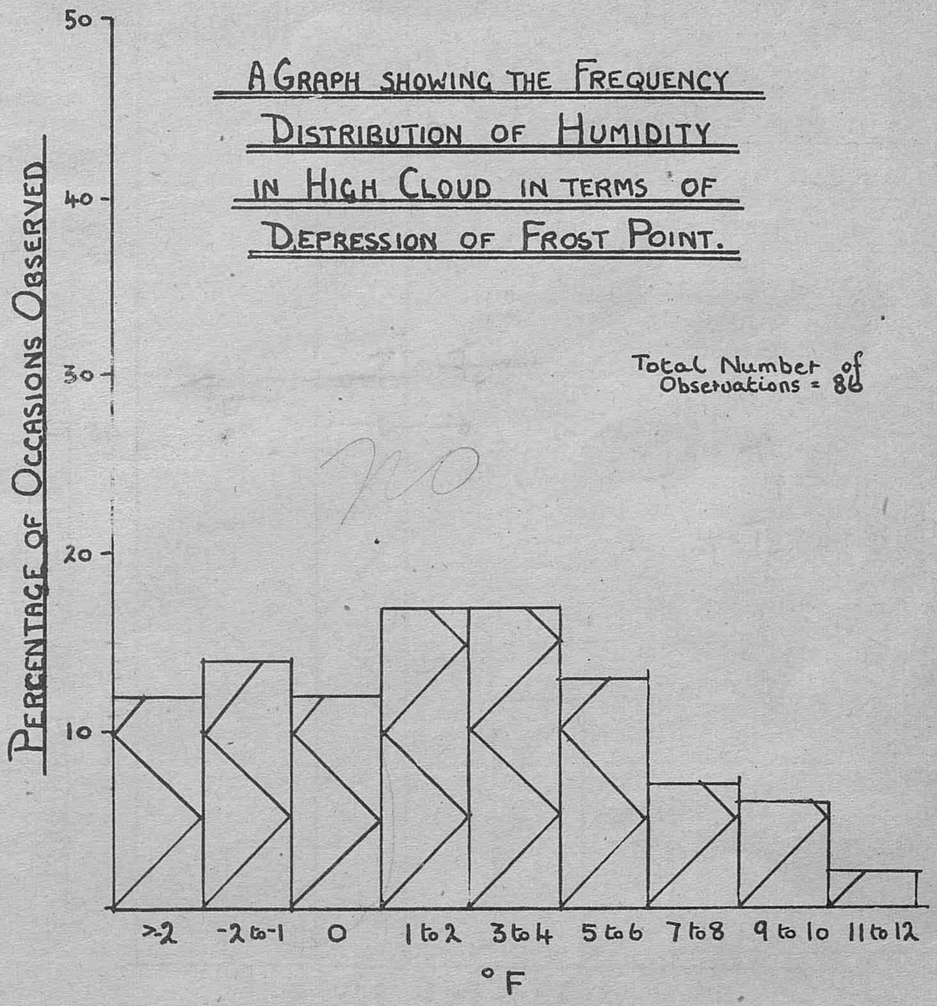


A Graph showing the Frequency Distribution of the Thickness of High Cloud

Total Number of Observations = 63

Thickness of Cirrus Cloud

Figure 5.



A Graph showing the Frequency Distribution of Humidity in High Cloud in terms of Depression of Frost Point.

Total Number of Observations = 86

no

Minimum Depression of Frost Point measured in Cirrus Cloud

Figure 9.

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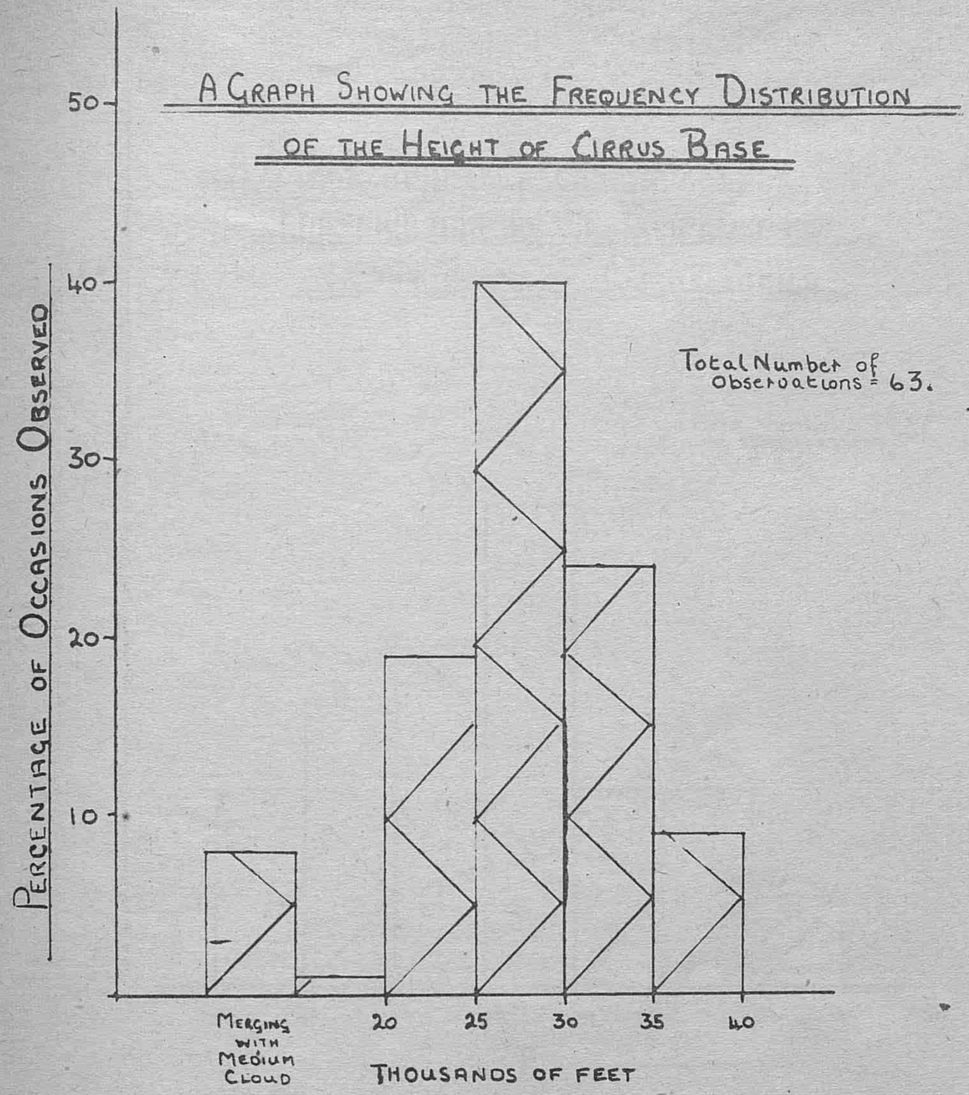
## A GRAPH SHOWING THE FREQUENCY DISTRIBUTION OF THE HEIGHT OF CIRRUS BASE

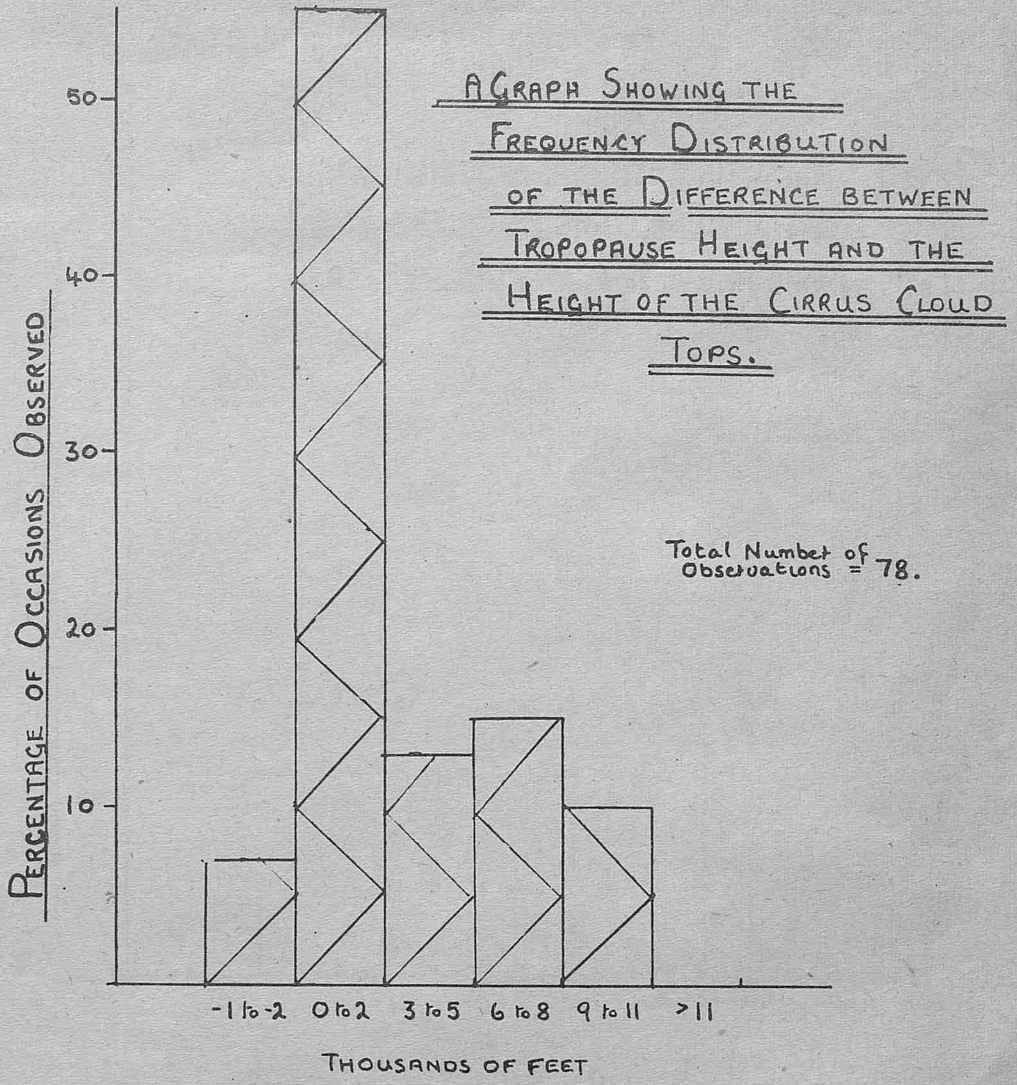


THE HEIGHT OF THE CIRRUS BASE.

FIGURE 3.

Fig 3

## A GRAPH SHOWING THE FREQUENCY DISTRIBUTION OF THE DIFFERENCE BETWEEN TROPOPAUSE HEIGHT AND THE HEIGHT OF THE CIRRUS CLOUD TOPS.



THE DIFFERENCE BETWEEN THE HEIGHT OF THE TROPOPAUSE AND THE HEIGHT OF THE CIRRUS CLOUD TOPS.

FIGURE 4.

Fig. 15

M.R.P. 833.

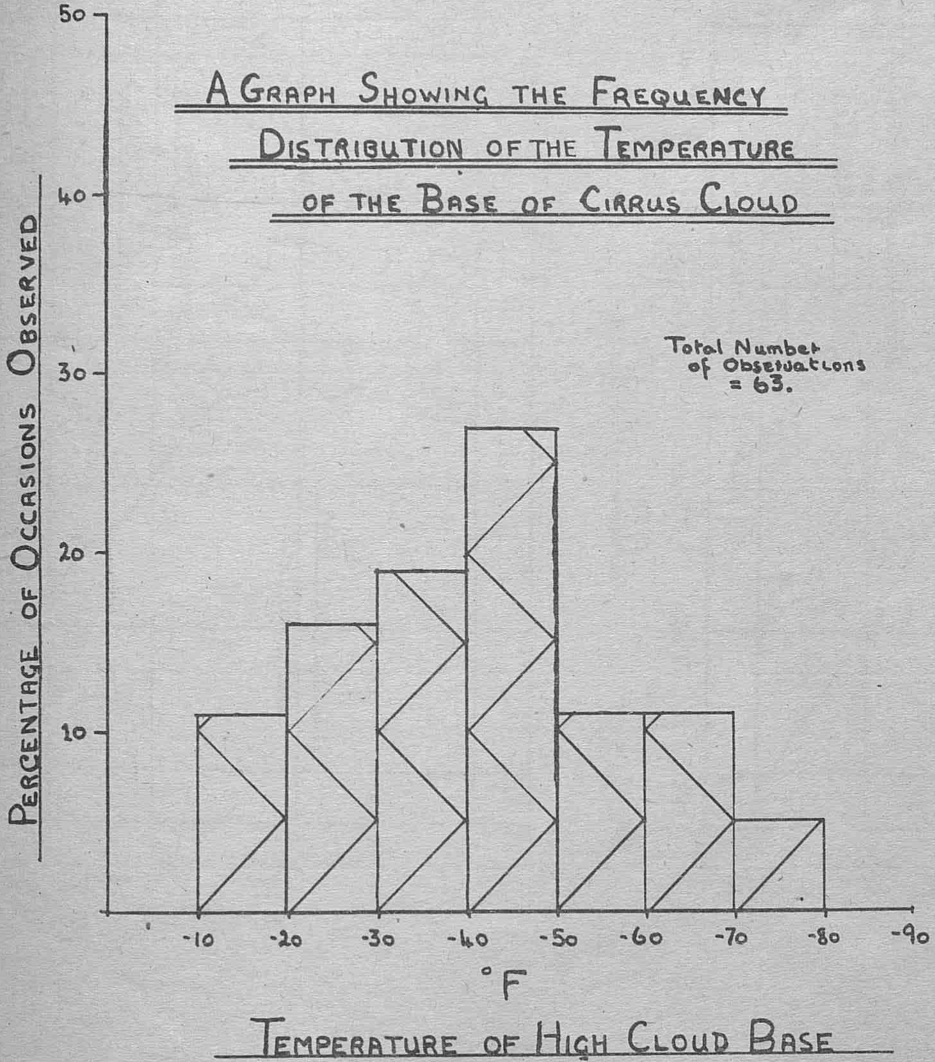


Figure 7.

Fig 12

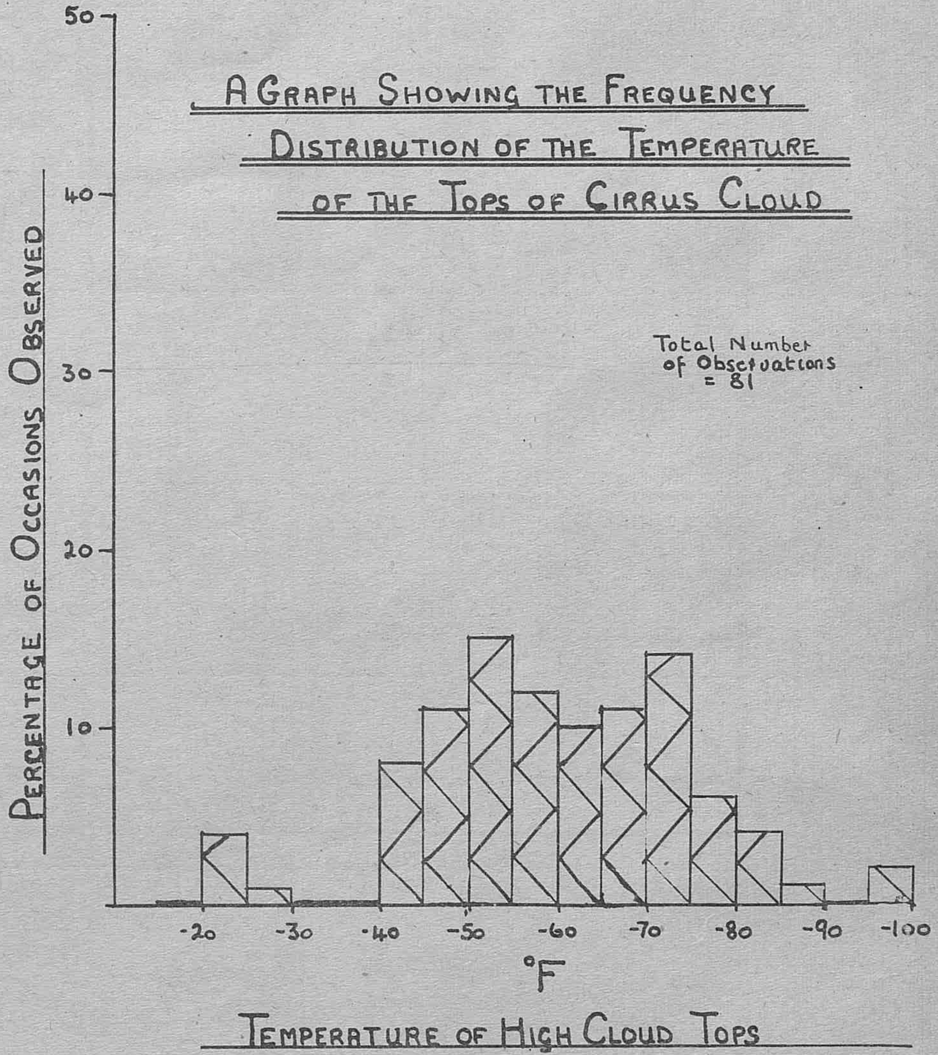


Figure 8.

Fig 13

# THREE AIRCRAFT SOUNDINGS, ILLUSTRATING THE "MOIST LAYER" EFFECT

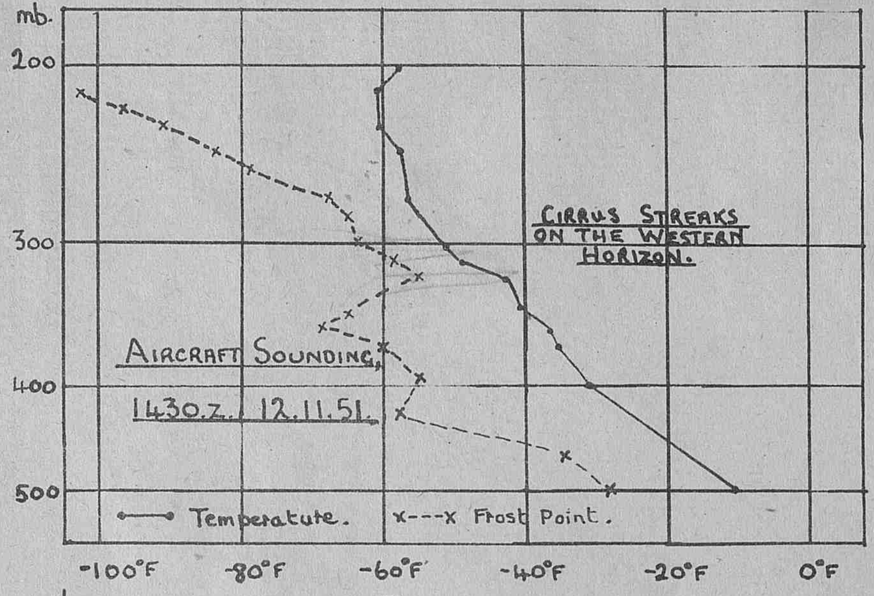


FIGURE 10.A.

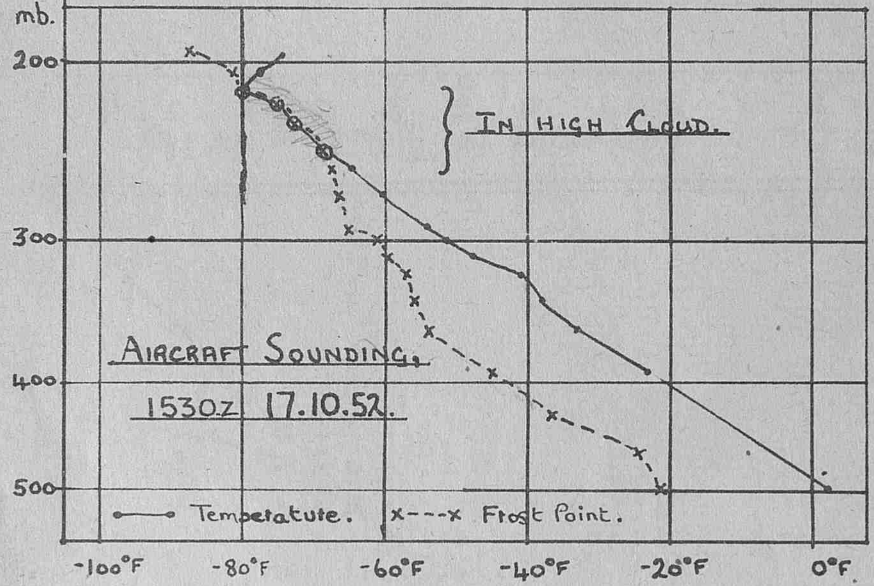


FIGURE 10.B.

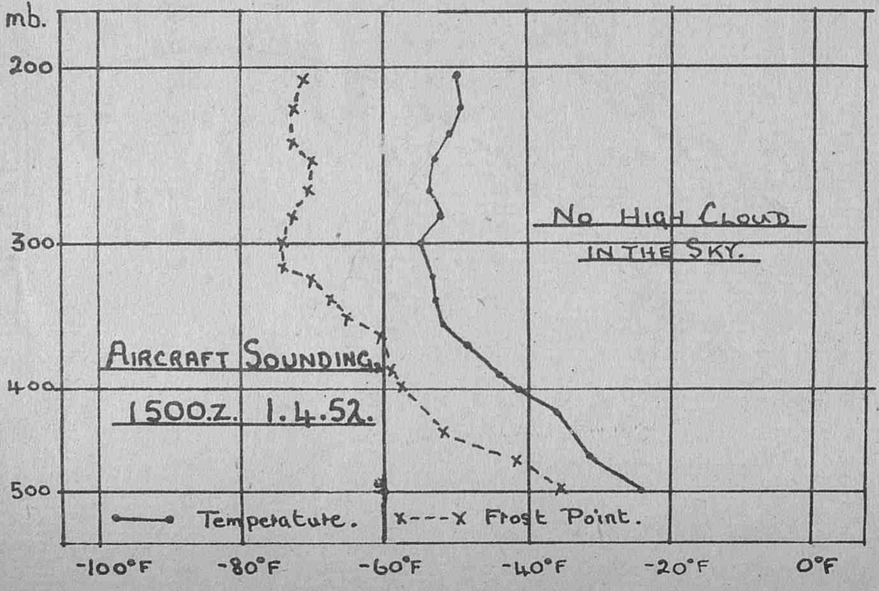


FIGURE 10.C.

A GRAPH OF THE HEIGHTS OF THE MOIST LAYER BASE AND TOPS PLOTTED AGAINST THE HEIGHT OF THE TROPOPAUSE.

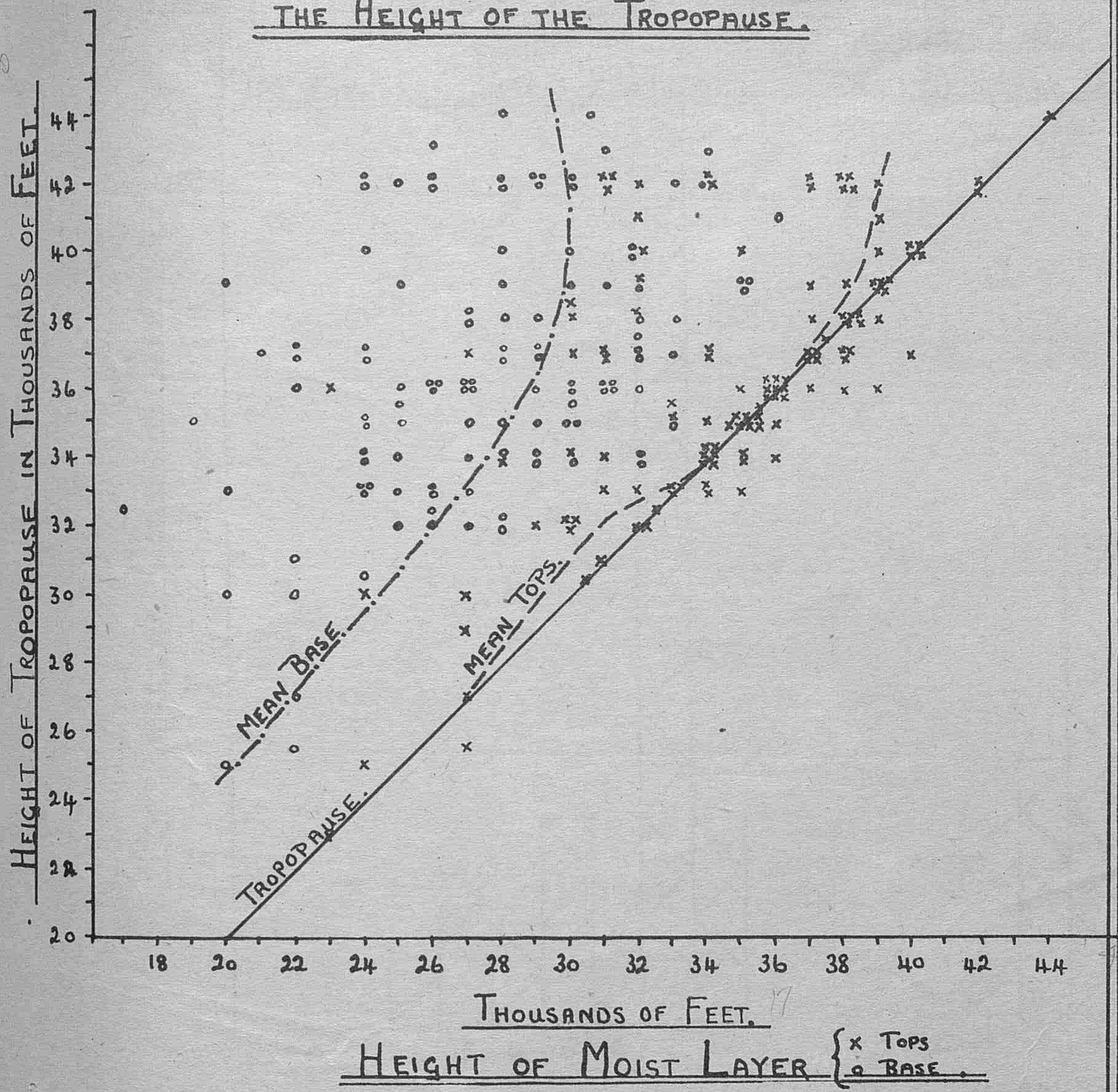


FIGURE 11.

Fig. 22

A GRAPH OF THE HEIGHTS OF HIGH CLOUD BASE AND TOPS PLOTTED AGAINST THE HEIGHT OF THE TROPOPAUSE.

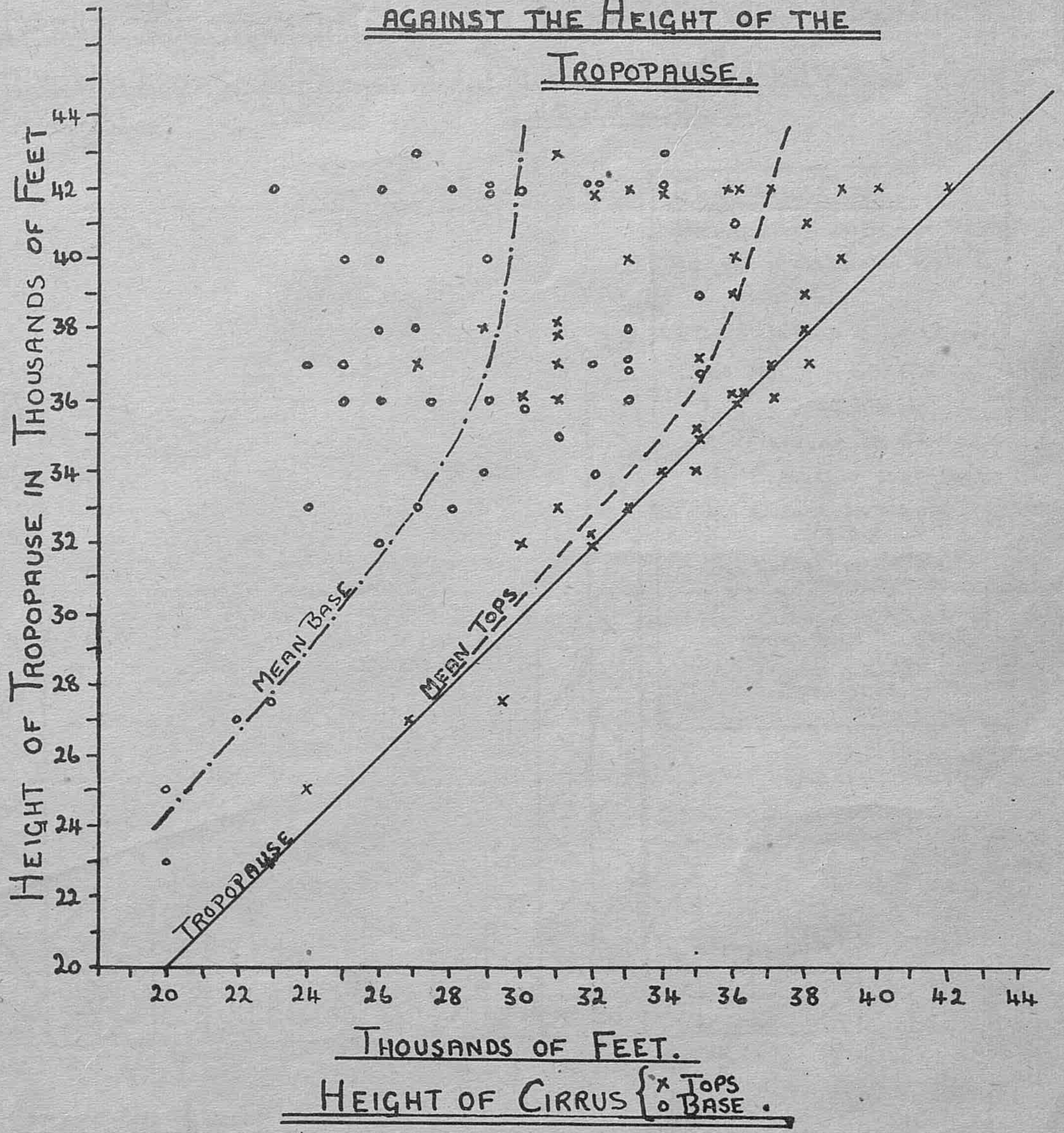


FIGURE 6.

Fig. 19

M.R.P. 833.

GRAPHS SHOWING THE FREQUENCY DISTRIBUTION OF THE DEGREE OF PERSISTENCE OF CONTRAILS OBSERVED ON FLIGHTS WHEN:–

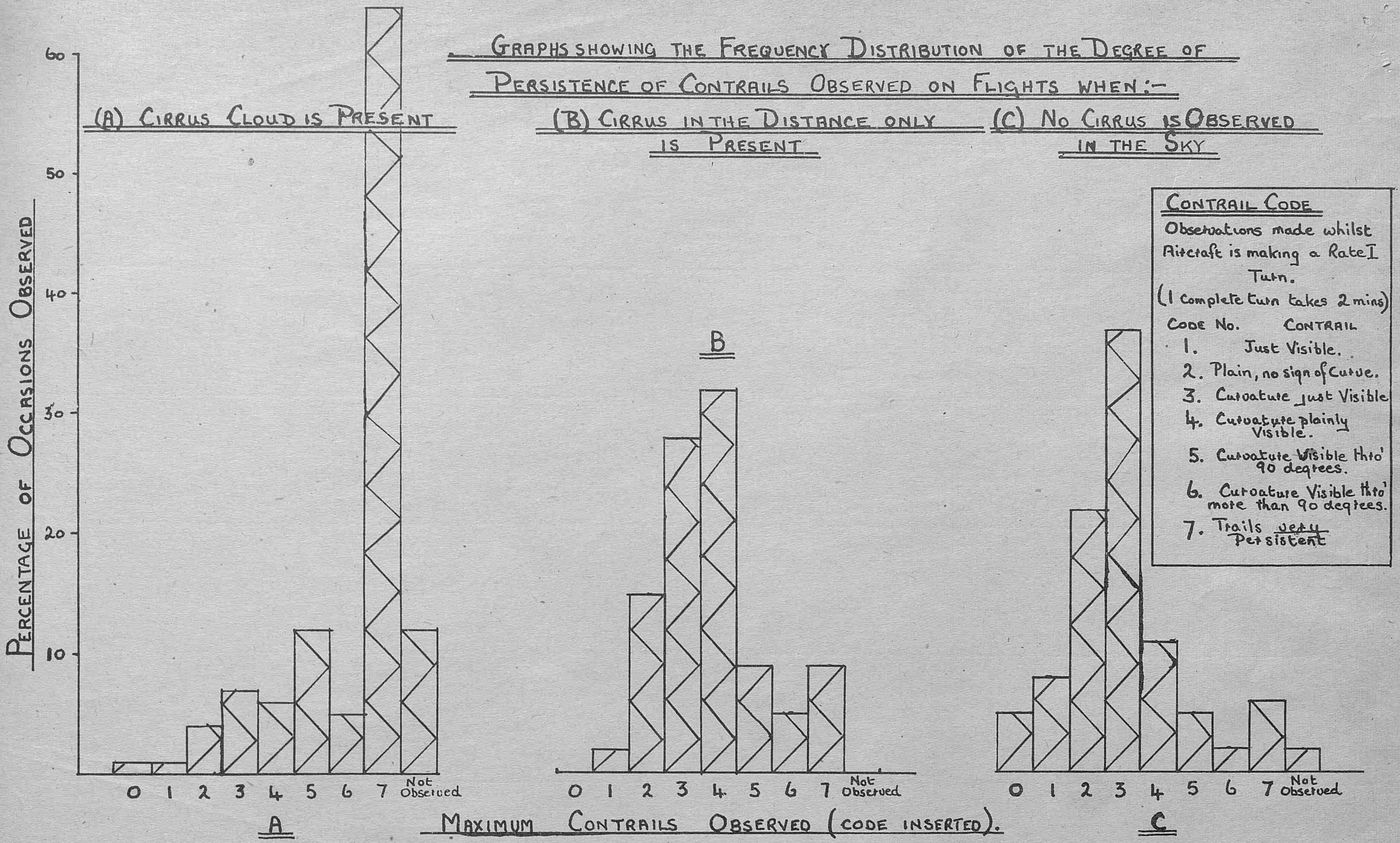


FIGURE 12.

MRP's 834, 836, 837

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London),
M.R.P. No. 834 by Mr. H. H. Lamb
A copy is available in the Library of the Meteorological Office."

M.R.P. 834

S.C.II/159

29th September, 1953

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METEOROLOGICAL RESEARCH COMMITTEE

# Two-way relationships between the snow/ice limit and 1000-500 mb. thicknesses in the overlying atmosphere

by

H.H. Lamb (Meteorological Office)

## Summary

Statistical studies of 1000-500 mb. thickness (i) with frozen or unfrozen forms of precipitation occurring and (ii) respectively over regions of extensive snow/ice - covered surface, over bare ground and open sea are used to discover relationships between the thickness values prevailing and (i) the likelihood of rain or snow falling (ii) the snow/ice limit on the ground. The results indicate an important influence produced by snowfields of the order of 1,000 miles and more in longitudinal extent upon the general circulation of the atmosphere.

1. Introductory. 1000-500 mb. thickness maps (relative topography) and the associated 500 mb. contours (absolute topography of the 500 mb. surface) have become the main tools of the busy forecaster concerned with co-ordination of much upper air data and of the research analyst studying broad-scale weather patterns. The 500 mb. level is the highest for which enough reliable data are available for analysis over wide areas, and these charts provide a good synopsis for study of the basic thermal and dynamical problems.

The investigation here described was undertaken firstly to establish a readily usable criterion for the forecaster with regard to thickness values and the probability of rain or snow falling, and secondly to probe in more detail an evident association between the 17,400 ft. thickness isopleth and the contemporary limit of snow-cover and of fast ice on the polar seas.

The results suggest that a widespread snow-cover behaves as an important control reacting upon the atmospheric circulation itself.

2. Vertical temperature distribution and 1000-500 mb. thickness. To be able to follow the working of climatic controls we commonly focus our attention on particular thickness isopleths, conveniently marked by standard colours on the working charts, as introduced by Dr. Sutcliffe. Details are given in Table 1.

Table 1. Standard thickness isopleths and corresponding temperatures.

| 1000-500 mb. Thickness | | Corresponding Temp. at 1000mb. (for saturated Adiabatic Lapse Rate throughout) | | Standard Colour of Isopleths |
|---|---|---|---|---|
| Feet | Metres | °C. | °F. | |
| 19,200 | 5,850 | 27 | 81 | Purple |
| 18,600 | 5,670 | 20 | 68 | Red |
| 18,000 | 5,490 | 13 | 55 | Green |
| 17,400 | 5,310 | 6 | 43 | Blue |
| 16,800 | 5,130 | - 1.5 | 29 | Brown |
| 16,200 | 4,940 | - 9 | 16 | Purple |
| 15,600 | 4,760 | -16.5 | 2 | Red |

Fig. 1 presents the distribution of temperatures aloft corresponding to these particular thickness values in certain circumstances:

(i) Full lines - saturated adiabatic lapse rate throughout the 1000-500 mb. layer.

(ii) Pecked lines - dry adiabatic lapse rate from 1000 mb. up to the 850 mb. level, saturated adiabatic lapse rate from 850 to 500 mb.

It will be seen by eye that strong inversions of temperature confined to the surface layer (often with pressures over 1000 mb) must have little effect on the temperatures above 900 or 950 mb. corresponding to particular thickness values. Larger positive temperature departures in the upper levels from the curves shown in fig. 1 are only likely to occur with deep inversions or isothermal layers - i.e. with a few fronts and occasionally in other situations in continental interior or inner Arctic regions.

In warm climates, where the air may have a high moisture content, temperatures indicated by the curves of state in fig. 1 should be virtual temperatures, exceeding the dry bulb temperature by up to 5°C (9°F) - the figure for saturated air at an observed temperature of 30°C (86°F.). At normal surface temperatures in middle latitudes the correction for virtual temperature is seldom more than 1 to 2°C., and at the freezing point or below the discrepancy is quite negligible.

In order to form a more realistic gauge of the dry-bulb temperatures actually occurring with the particular 1000-500 mb. thicknesses corresponding to our standard lines, all available upper air ascents from stations in western, central and northern Europe and from the weather ships in the eastern north Atlantic for the year 1951 which gave thickness values of just 16,800; 17,400; 18,000 or 18,600 within ± 30 ft. were statistically examined.* The results of this examination are set forth in Table 2.

To reduce the work to manageable proportions, the 18,000 ft. thickness cases studied were limited to those occurring in July, November and December.

/Table 2.

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* FOOTNOTE. The stations examined were:-

British ocean weather ships near 59°N 19°W and 52°N 20°W

Thorshavn, Lerwick, Stornoway, Leuchars, Aldergrove, Valentia, Liverpool, Larkhill, Camborne, Downham Market, Sola, Gardermoen, Torslanda, Stockholm, Riga, Warsaw, Copenhagen (Kastrup), Flensburg, Jever, Iserlohn, Hanover, Berlin (Tempelhof), Greifswald, Wernigerode, Dresden, Erlangen, Munich (surface temperatures not used: altitude 1,700 ft.), Wiesbaden, Utrecht (De Bilt), Brussels (Uccle), Paris (Trappes), Strasbourg, Bordeaux, Lyons, Nimes, Payerne (surface temperatures not used: altitude 1,500 ft.), Vienna.

Checking weeded out some doubtful ascents. In the end 293 ascents from these 39 places were used and are summarized in Table 2.

## Table 2. Observed Temperatures °F. with particular 1000-500 mb. Thickness values.

| 1000-500 mb. Thickness ft. | Surface Temps. °F. Mean | S.D. | Extremes | Temps. at 850 mb. °F. Mean | S.D. | Extremes | Temps. at 700 mb. °F. Mean | S.D. | Extremes | Temps. at 500 mb. °F. Mean | S.D. | Extremes | No. of Cases |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 16,800 | 25.7 | 10.3 | 37<br>-1 | 11.3<br>-11.5 | 3.8 | -8.3<br>+17<br>+ 5<br>-15.0 | -20.8<br>-5.5 | 2.4 | -18.9<br>-2<br>-9<br>-22.8 | -36.9<br>-38.3 | 5.4 | -27 -32.8<br>-44 -42.2 | 18 |
| 17,400 | 40.1 | 5.6 | 52<br>28 | -3.6<br>25.5 | 2.9 | +0.6<br>33<br>18<br>7.8 | 12.4<br>+9.6 | 2.3 | +16 8.9<br>+ 3 16.1 | -18.8<br>-28.2 | 4.2 | -23.3<br>-10<br>-28<br>-33.3 | 160 |
| 18,000 | 51.6 | 6.2 | 63<br>33 | +3.7<br>38.7 | 2.9 | 6.7<br>44<br>32<br>0 | 2.8<br>27.0 | 2.1 | 31 -0.6<br>21 6.1 | - 1.1<br>-18.4 | 2.6 | + 4 15.6<br>- 6(July, Nov., Dec. only)<br>-21.1 | 41 |
| 18,600 | 69.7 | 7.6 | 84<br>50 | 13.0<br>55.5 | 3.1 | 17.8<br>64<br>49<br>9.4 | 4.3<br>39.7 | 2.3 | 7.2<br>45<br>34 1.1 | -10.7<br>+12.8 | 3.2 | -5.0<br>+23<br>+ 5 -15.0 | 74 |

The striking feature of Table 2 is that standard deviations are only about 3°F. at all the upper levels, save for the top of the 1000-500 mb. layer in the coldest ascents. The actual mean temperatures are close to the standard curves in fig. 1.

Many different groupings of the ascents according to time of day, time of year and geographical position were tried. The resulting mean temperatures differed in nearly all cases by less than 2°F. at 500 and 700 mb. and, in most cases, also at 850 mb. from the overall means in Table 2. Greater differences occurred at the ground, but these were entirely obvious. Only the following selected cases (Table 3) were of any special interest.

## Table 3. Observed Temperatures °F. in selected situations.

| 1000-500 mb. Thickness ft. | Mean Temp. at Ground °F. | Mean Temp. at 850mb. °F. | Mean Temp. at 700mb. °F. | Mean Temp. at 500mb. °F. | No. of Cases | Basis of selection | |
|---|---|---|---|---|---|---|---|
| 16,800 | 18.3 | 9.1 | - 4.7 | - 33.4 | 9 | Continental group | Stockholm<br>Torslanda<br>Oslo |
| 16,800 | 31.0 | 13.4 | - 6.4 | - 40.4 | 9 | Oceanic Group | Ships<br>Thorshavn<br>Lerwick<br>Stornoway |
| 17,400 | 44.2 | 25.0 | + 8.6 | - 17.6 | 22 | Weather ships | (all cases) |
| 18,600 | 62.0 | 55.0 | 41.9 | + 15.5 | 8 | Oceanic Group | Ships<br>Valentia<br>Camborne |

Comparison of the cases in Table 3 with Table 2 brings out chiefly the great difference of behaviour of surface temperature over the ocean and at inland stations including in the British Isles. The greatest standard deviation was in surface temperatures at inland stations with 16,800ft. thickness, namely 11.2°F., whereas the S.D. of surface temperature at oceanic stations was under 3°F. alike with thicknesses of 16,800 and 18,600 ft.

Table 4 presents a survey of the behaviour of the freezing level in all the cases covered by Table 2.

/Table 4.

Table 4. Observed freezing levels with particular 1000 - 500 mb. thickness values.

| 1000-500 mb. Thickness ft. | Average level of (lowest) 0°C.Isotherm | S.D. | No. of Cases | Lowest observed 0°C. isotherm | Highest observed 0°C. isotherm | Percentage of all cases examined which showed double or triple freezing levels. |
|---|---|---|---|---|---|---|
| 16,880 | Surface | - | 18 | Surface | 960 mb. | 0 |
| 17,400 | 915 mb. | 37mb. | 160 | Surface | 850 mb. | 4.4 |
| 18,000 | 770 mb. | 34mb. | 41 | 860 mb. | 710 mb. | 12.2 |
| 18,600 | 635 mb. | 21mb. | 74 | 670 mb. | 570 mb. | 2.7 |

Summarizing, we find that the thickness value alone will give us the temperature at any given height in the free air to within 3°F., and the freezing level to within about a thousand feet, in a considerable majority of cases over the European and Atlantic region.

3. The freezing level criterion for rain or snow. The height of the freezing level above the ground and the thermal structure of the lowest few thousand feet of the atmosphere largely determine whether snow can reach the ground. A brief survey of this point is essential before proceeding further.

R. Murray (1952) examined the occurrence of rain (and rain showers) or snow (and snow showers) with various freezing level heights in the British Isles and found that frozen and unfrozen forms become equally likely when the freezing level is about 1,000 feet. This figure seems quite generally applicable to continental and oceanic locations.

The problem is however complicated somewhat for the forecaster by time-variations of freezing level within one air-mass. Quick, erratic variations of several hundred feet occur in instability showers; these are usually fleeting, but under favourable circumstances a new snow carpet may result from showers starting with a general freezing level above 1,000 feet. Douglas (1950) noticed that long-continued rain and sleet falling through dry air may so cool the atmosphere that the freezing level drops by as much as 4,000 to 5,000 feet (forming a deep isothermal layer about 0°C.) and snow ultimately reaches the ground. Widespread snowfalls in southern England began in this way on 26.12.1927 and 1.11.1942; the same effect is both frequent and important in central Europe where low relative humidities favouring evaporation cooling are much commoner than in England.

These considerations come into play with 1000-500mb. thicknesses from 17,100 feet up to 17,500 or, at most, 17,600 ft. in connexion with snowfall over regions of general elevation less than 100 to 150 metres (300 to 500 ft.).

4. The thickness criterion for snowfall: precipitation in frozen and unfrozen forms. R. Murray (loc. cit.) statistically examined the occurrence of precipitation in frozen and unfrozen forms with various thicknesses of the 1000-500 and 1000-700mb. layers at or near seven British upper air stations (Lerwick, Stornoway, Aldergrove, Leuchars, Liverpool, Downham Market and Camborne). Grouping these stations together inevitably masked the differentiation we have noticed between inland/continental and oceanic localities, most of his chosen stations being classable as oceanic. Murray found that rain and snow were equally likely when the 1000-500 mb. thickness was about 17,140 ft. or the 1000-700 mb. thickness about 9,120 ft.: these figures prove to be fairly typical of oceanic (Gulf Stream water) situations. Evidence of generally rather higher thickness values for the changeover from rain to snow may be detected in a recent paper of the Meteorological Research Committee on cold pools by Sumner (1952), who found a "fairly distinct separation" when the minimum thickness in the cold pool was about 17,200 feet.

To establish the most general possible criterion, based upon a broader survey, the present investigator tabulated all cases of precipitation reported during the winter 1950/51 at the time (including precipitation in the last hour) and in the neighbourhood of the radio-soundings at Lerwick, Aldergrove, Liverpool (surface observations from Manchester, Ringway), Downham Market (Mildenhall), Berlin (Templehof), Riga, Stockholm and Oslo (Fornebu). These are all low-level stations, less than 300 feet above mean sea level; they comprise one oceanic location (Lerwick), various inland districts of the British Isles and several typical European continental locations. The mean latitude of the stations used is 56°N. The months examined were November 1950 to April 1951 inclusive for the British stations and October 1950 to April 1951 inclusive for the others. In addition the precipitation forms at one higher-level station in the British Isles, Eskdalemuir (794 ft. A.S.L.), were studied in relation to the contemporary 1000-500 mb. thickness values over the station taken from the published Daily Aerological Record.

In all there were 2,045 observations with precipitation to be classified.

The separate frequencies of rain, drizzle, snow, sleet, hail and showers of each type, also the total frequencies of precipitation of showery character, occurring with thickness ranges 17,960-18,050, 17,860-17,950 and so on down to 16,860-16,950 feet were examined for each of the nine stations named and in each month of the winter. Some of the results are shown in the histograms reproduced in figs. 2(a)-(f), in which sleet and hail are counted in with the frozen forms and shower frequencies include all forms of showers. Results for the whole collective and the more important variations may be summarized as follows:-

(a) Rain or frozen forms. In general the changeover was commonest with thickness about 17,300 feet, at which figure frozen and unfrozen forms were equally frequent.

Over the warm waters of the North Atlantic Drift as at Shetland (water surface temperatures in winter about 10°C.(50°F.)) and the coastal places in the British Isles studied by Murray, the changeover occurs about 17,150 feet. In such situations snow only falls in air which is unstable (or nearly so) from the surface up.

Over already extensively snow-covered terrain, as at Oslo, Stockholm and Riga in the depth of winter, snow is as likely to fall as rain with thicknesses 17,400 to 17,450 ft. or even higher.

The critical thickness at inland stations in the British Isles and at Berlin approximated either to values typical for Oslo-Stockholm-Riga or to the Atlantic group according to whether there was extensive snow lying or not.

Snow (or sleet) which fell with the higher thickness values, and all cases of snow with thicknesses over 17,700 feet, were associated with considerable inversions of temperature or deep isothermal layers and the precipitation may be presumed to have originated in a stratus or stratocumulus cloud-sheet below the stable layer. It is thought that in all cases with thicknesses of 17,700 feet or more, the ground was already snow-covered. Cases which appeared to substantiate this as a general rule were noted in such unlike regions as Aldergrove (N. Ireland) and Oslo.

Data were insufficient for any thorough examination of variation with height of the liability for snow rather than rain with different 1000-500 mb. thickness values. The results for Eskdalemuir (about 800 feet above sea-level) gave a more "blurred" criterion than for the low-lying stations examined and may have been as much affected by frequent occurrence of widespread snow lying over the Southern Uplands region in which the station lies as by its altitude.

/(b)

(b) Sleet. Sleet was common only over a small range of thicknesses around the for-the-time-being critical value for rain or snow. A few occurrences of sleet were noted with thicknesses as low as 17,100 feet at both Lerwick and Stockholm. Sleet may also occur with thicknesses up to 17,700 feet; but in the higher thickness ranges a clear-cut division between snow and rain is more usual, depending on whether the precipitation originates below the inversion or above it.

(c) Hail. Hail was rare or unobserved in winter at all the inland and continental stations. When it did occur it was only small hail, falling with any thickness values below 17,600 feet. On the other hand the larger forms of hail, such as occur in the Mediterranean area and Atlantic coasts in winter and inland in the British Isles in summer, are known to occur with 1000-500mb. thicknesses up to 18,000 feet or above.

(d) Drizzle (including supercooled or freezing drizzle). Drizzle was observed with 1000-500 mb. thicknesses as low as 17,000 feet in the British Isles and 17,100 feet in the Baltic region. Presumably drizzle may occur at any temperature which is not too low to permit enough supercooled droplets of suitable size to form in low cloud or fog. The relative frequency of drizzle increased with increasing thickness (cf. fig. 2(a)).

(e) Showers. At Lerwick, Aldergrove and Eskdalemuir, where the influence of the Gulf Stream water is strongest, showers became more frequent the lower the thickness. At continental and inland stations the maximum frequency of showers occurred with thicknesses about 17,300 feet. Lower thickness values were mostly associated with continental air which was either too dry or too stable in the lower layers to give showery precipitation.

Table 5 gives a survey of the results for all the stations examined for the winter period December 1950 - February 1951 and shows the main groupings.

Table 5. 1000-500mb. thicknesses (in feet) accompanying rain or snowfall at various stations (December - February)

| Station | Value giving equal likelihood of rain or snow | Highest thickness with snow | Lowest thickness with rain | Remarks |
|---|---|---|---|---|
| Lerwick | 17,150 | 17,400 | 17,100 | Sharp changeover |
| Eskdalemuir | 17,550 | 17,700 | 17,100 | No sharp criterion |
| Aldergrove | 17,300 | 17,700 | 17,100 | Fairly sharp changeover |
| Manchester | 17,300 | 17,600 | 17,200 | - ditto - |
| Mildenhall | 17,300 | 17,400 | 17,200 | - ditto - |
| Berlin | 17,300 | 17,700 | 17,100 | - ditto - |
| Oslo | 17,400 | 17,700 | (17,300) | Less data than for above stations |
| Stockholm | 17,400 | 17,700 | 17,200 | Sharp changeover |
| Riga | 17,600 | 17,700 | (17,300) | Data insufficient |

(f) Seasonal differences. Monthly data for individual stations were generally too scanty to reveal anything significant. Various groupings, such as the monthly trends of critical thickness values at oceanic and continental stations, were looked at. No seasonal trend emerged, apart from an evident requirement of really low thicknesses (17,200 ft. or below) for late spring snow and sleet in the British Isles from about mid-April onwards. This observation is doubtless related to rising sea-surface temperatures and increasing remoteness of the sources of cold air. Exceptionally early snowfall in autumn might be expected to demand similar conditions.

/(g)

(g) Geographical range. The main differentiation noted under (a) was between oceanic and continental locations and as to whether a snow-cover already existed or not. Subject to this distinction, the criteria for rain or snow (but not for hail) on low ground should have world-wide application.

As a check, 1000-500 mb. thicknesses over Malta, lying outside the normal limit of snowfall at the southern fringe of Europe, during winter 1952/53 were tabulated. The island is small and the surrounding water has a surface temperature in winter of about 13°C.(55°F.). The lowest thicknesses observed, 17,250 to 17,500 feet, were associated with cold rain and on several days in January with wet sleet falling at points over 200 feet above sea level. Snow was not observed. Hail with stones up to 1 cm. diameter fell with thicknesses ranging from 17,500 to 18,000 feet.

5. Association between the limit of an established snow or ice surface and 1000-500 mb. thicknesses in the atmosphere. A snow or ice surface has a very different effect upon the air over it from that of either dry or moist land or deep water. The diurnal range of temperature of the actual surface, whether of snow or of extensive sea-ice, is large and of the order of that of dry ground. The surface radiates out heat approximately as a black body, but a high proportion of incoming radiation is reflected and lost. The maximum temperature of the surface cannot in any case rise above 0°C. until the snow or ice is removed, a process delayed by the requirement of much latent heat to be supplied from the air, from warm rain, from the underlying soil (or water) or by direct insolation.

Studies of ablation suggest that invasions of mild air over extensive snowfields in northern Europe in winter, typically accompanied by screen temperatures of the order of +3°C.(37°F.) at a height of 4 feet above the snow, normally thaw about 1 inch of snow in 24 hours, if there is no rain or drizzle; the ablation increases to about 2 to 4 inches in a day if accompanied by appreciable rainfall.* Disappearance of 6 or 8 inches of snow may be expected to require either a continuous supply of mild air of several days' duration or a fall of an inch of rain or more.

Composite maps showing the position of the snow-limit on successive days indicate that this limit behaves in such a way that rapid advances, often on a wide front, are followed by periods of stability or of more gradual retreat in which certain segments only of the snow margin are lost at any one time.

The rapid advances result from passage of a single depression or invasion of a single cold airmass giving instability showers. The process described in an earlier section whereby long-continued precipitation chills air that was formerly rather mild and dry is also liable to result in important advances of the snow margin over previously bare ground.

Withdrawals of the snow margin follow all types of ablation. In central and northern Europe, including the British Isles, warm rain is the most effective agent for removing snow-cover in winter. At other seasons, and in regions south of about 45°N even in winter, sunshine is most effective.

Positions of the snow limit which coincide more or less with a mountain barrier are more liable to persist, when warm air advances over the high ground (and commonly fails to descend to the low ground on the lee side), than positions anywhere over the European lowlands where no barrier exists. The effect is sufficiently obvious in the way the course of the mean 0°C. surface isotherm over Europe follows the topography, with several successive weekly or monthly positions remaining close to the Norwegian and central European massifs. The greater frequency of snow lying everywhere over northern and eastern England east of the main watershed compared with the same levels on the western side (Manley 1947) is also at least partly attributable to this cause.

/Extensive

Footnote * See for instance Champion (1950 and 1951).

Extensive snow lying cools warm air passing over it in a largely similar manner to the case of warm air in summer over the Arctic ocean and other cold-water regions (downward eddy-flux of heat) (cf. Taylor 1914, Sverdrup 1936, Scherhag 1937). Over winter snow there may be, additionally, important radiation cooling from the top of fog and stratus formed in the mild air. Where the snow limit coincides with a mountain barrier, any forward thrust of warm air may be lifted above the surface and is liable to contribute frontal and orographic snowfall which adds to the depth of the lying snow.

These effects modify the air and limit the freedom of movement of the thickness isopleths.

The writer therefore decided to investigate the occurrence of various 1000-500 mb. thicknesses over established snow and ice and in the neighbourhood of the snow limit.

Positions of the snow limit over Britain and Germany were obtained by mapping actual state-of-ground reports. The number of cases in which the snow-limit was accurately known in this way was augmented, for the purpose of the tests, by many other cases in which the position of the limit could be deduced beyond reasonable doubt, to within 100 miles, from consideration of the reported sequence of weather for a week or more and the behaviour of the surface air temperatures. The last-named are particularly revealing whenever a broad warm airstream encounters lying snow and a stationary belt of sharp fall of surface temperature to +3°C. (37°F.) or lower in the moving airstream marks the snow limit. This reasoning frequently made it possible to place the snow margin in many parts of Europe and N. America. Composite maps were then drawn showing successive daily positions of the limit of snow and ice.

(a) Thickness values at the limit of snow - and ice - covered surface. For this test reported 1000-500 mb. thicknesses and values estimated from the isopleths were taken at points over the western and southwestern limit of contemporary snow-cover in Europe between 45 and 60°N, care being taken to choose points where the snow edge was fairly straight and free from deep embayments. Points used were in west Norway, western Britain, France, Germany, Poland or Sweden in a region well covered by upper air observations. Over 80% of the points chosen were on ground less than 300 feet (100 metres) above sea level, the remainder being at the edge of a mountain barrier; in neither case did it appear necessary (either *a priori* or in the light of results obtained) to make any allowance for the elevation of the actual test point, so long as all the low ground (below 300 ft.) to leeward was snow-covered. The region studied was also entirely within the zone of prevailing westerly winds aloft: any association found between the thickness value and the windward limit of snow-covered land (in the sense of the prevailing upper westerlies) would therefore be liable to imply some distortion of the atmospheric circulation aloft.

The mean value of 1000-500 mb. thickness given by 1,053 observations over the best attested positions of the snow limit in this European region on 99 different days between January 1947 and February 1952 was 17,297 feet, the standard deviation being about 170 feet.

This figure should be compared with normal thicknesses between January and early March in different parts of the region studied ranging from about 17,200 to 17,700 feet with a S.D. at any given place of the order of 300 feet. (e.g. mean thicknesses for Larkhill for two successive Januaries were respectively 17,460 feet S.D. 523 feet and 17,657 feet S.D. 259 feet.)

The low mean thickness and small standard deviation at the snow-limit, wherever that may happen to be, point to a real association between the snow limit and the 1000-500 mb. thickness over it and to a contemporary shift of the upper wind pattern (thermal and actual).

Various additional tests were made to broaden the basis of this result and to gain some indication of its application to other geographical regions.

The average 1000-500mb. thickness, from eighty-five radio-sonde ascents made by the Dutch whaling ship Willem Barendsz, near the pack-ice in 60°S between 12°E and 30°W (December 1946 - April 1947) in the South Atlantic was 17,260 feet S.D. 212 feet. The standard deviation might reasonably be supposed to be a little less if taken precisely at the ice margin in all cases.

As no further actual positions of the snow or ice limit were readily available, tests were made of the mean 1000-500 mb. thickness values at points on the normal position of the M.S.L. 0°C. isotherm shown by the official Russian Climatological Atlas (published 1930) for Russia and Siberia and by Bartholomew's Atlas of Meteorology (published Edinburgh 1899) for North America. The Russian and Siberian material supplied mean positions for the 0°C. isotherm at about 10-day intervals from 11th September through the winter to 1st June. The North American material yielded only monthly mean positions, which were taken as applying to the middle day of each month. The means related to the years 1849-1899 for Siberia, 1881-1915 for European Russia and 1870-1884 for N. America. An obvious weakness was that the mean positions of the 0°C. isotherm at sea level for these years had perforce to be compared with mean 1000-500 mb. thicknesses about the same date of the year for the period 1946 to 1950. In practice thickness means for seven days centred upon the relevant date were taken at each intersection of the 0°C. isotherm with the meridians of longitude, every ten degrees of longitude from 120°W to 60°W and from 20°E to 140°E. This yielded 3,268 values. The mean was 17,507 feet and the standard deviation 280 feet.

Considering the crudeness of the test, this result agrees about as closely as we could expect with the values obtained for the actual snow limit on the actual day and year. The discrepancy is likely to be due chiefly to two causes:-

(i) The 0°C. mean isotherm need not by any means coincide with the snow limit for a given period, but seems likely to protrude over bare ground somewhat beyond the snow-covered region.

(ii) The snow-cover has probably been rather less extensive in recent years, apart from the winter 1946/47 in the European sector, than in the nineteenth century.

From these tests it seems fair to conclude that the mean 1000-500 mb. thickness over the contemporary windward limit of snow and fast ice on the polar seas is close to 17,300 feet in all parts of the world and that the variation of thickness over the snow margin is distinctly smaller than that at any place well away from the margin.

(b) Movement of warm air and incursions of various thickness isopleths over extensive snow and ice-fields. Incursions of particular isopleths of 1000-500 mb. thickness were measured in nautical miles along the path of the gradient wind. Choice was restricted to cases where the snow limit was stabilized in a given position for a period of seven days or more. Note was then made of the maximum incursion of a given thickness isopleth along the instantaneous line of surface gradient wind, the mean strength of the gradient wind along that line at the time in question, the nature of the terrain and type of weather situation.

The measurements of maximum encroachment of various thickness values above 17,300 ft. over a widespread snow- and ice-covered terrain are fully logged in Table 6. Brackets indicate extreme penetrations which were excessively short-lived. All cases which may have been complicated by subsidence (deduced from anticyclonic curvature of the isobars and appropriate surface weather reports) are separately listed. Thirty-eight 7-day periods taken from six winters between 1st January 1947 and 29th February 1952 provided the required conditions for test measurements in different parts of the northern hemisphere including Alaska, Canada, Europe and western Siberia*. In intermediate cases the distance of greatest encroachment of a given isopleth over snow has been rounded upwards to the next hundred miles, whereas gradient wind ($V_G$ in the table) has been rounded downwards to the next lower unit in tens of knots. This means that, in spite of the rather small number of cases available for study, there is an enhanced likelihood of an observation as logged in the table approximating to the greatest incursion possible. Results from the first five winters (30 weeks in all permitting 35 measurements) were first logged as a separate group and analyzed graphically to discover likely asymptotes (maximum incursions and minimum gradient winds); thereafter results for the sixth winter (8 weeks in 1951/52 permitting 12 measurements) were similarly treated and compared. The additional material from the sixth winter (about 30% of the original data considered) repeated the same pattern and pointed to substantially the same extremes.

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FOOTNOTE:

* The terrains covered were of course infinitely varied and difficult to summarize. Maximum encroachments, measured either over extensive lowland (e.g. the north European plain) or over an initial mountain barrier which also marked the contemporary snow limit (e.g. the Alps and Alpine foothills), were however found to give a reasonable gradation with wind speed (Tables 6 and 7) without making any allowance for the elevation of the country.

/TABLE 6.

TABLE 6. GREATEST ENCROACHMENTS OF VARIOUS 1000-500 MB. THICKNESSES OVER SNOW AND ICE SURFACES, MEASURED ALONG GRADIENT WIND.
(All individual cases)

Distance of greatest incursion over snow/ice surface in Nautical Miles

(Tabulated figures below are numbers of cases)

| GRADIENT WIND AT M.S.L.<br>MEAN STRENGTH ALONG PATH | 0 | 100 | 200 | 300 | 400 | 500 | 600 | 700 | 800 | 900 | 1000 | 1100 | 1200 | 1300 | 1400 | 1500 | 1600 | 1700 | 1800 | 1900 | 2000 | 2100 | 2200 | 2300 | 2400 | 2500 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 70 Kt. |  |  |  |  |  | 1 |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 60 |  |  | 1 |  |  |  |  |  |  | 2 |  |  |  |  |  | (1) |  |  |  |  |  |  |  |  |  |  |
| 50 |  |  |  | 1 |  |  |  |  |  | 1 |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 40 |  |  |  | 1 | 2 |  | 3 | 1<br>* | 2 |  |  | ( *) | (1) |  | * |  |  |  |  |  |  |  |  |  | ( *) (( *)) |  |
| 70 Kt. |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 60 | 4 |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 50 | 2 |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 40 | 7<br>3* |  |  |  |  |  | ((1*)) |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 30 | 3<br>2* |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 20 | 1<br>4* |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 10 |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
| 0 | 6<br>2* |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |

(36 cases in all)

INCURSIONS OF THE 18,000 FT. ISOPLETH
Each case the **greatest** incursion in 7 days.

NOTE: Cases complicated by subsidence, indicated by anticyclonic curvature, are shown with an asterisk.
( ) means maximum incursion maintained for less than 48 hours.
(( )) means maximum incursion lasted under 12 hours.

We conclude from this table that:

(i) Thicknesses of 18,000 feet do not transgress at all over extensive snowfields without the aid of subsidence*.

(ii) Thicknesses of 17,800 ft. do not encroach more than about 500 nautical miles over snow without the aid of subsidence.

(iii) For thicknesses of 17,400 feet the maximum incursion is some 1,500 nautical miles without subsidence, but may reach 2,500 miles or more for a short time when subsidence is occurring.

In all cases the final limitation on greater incursions appears to be imposed by the improbability of gradient winds of the requisite strength being maintained over the necessary distances.

Table 7 lists the maximum encroachment* of the various thickness isopleths indicated by graphical analysis of the cases in the Table 6. (Figures in brackets are rather free estimates; hyphens have been entered where no estimate was justified by the data available.) The effect of subsidence or no subsidence may be clearly seen. A third column shows the result of singling out cases where the snow limit was stabilized near a mountain barrier athwart the path of invading warm air.

Coincidence of the snow limit with an orographic barrier (which need not be more than 1000 metres in height) may have a decisive effect. The right-hand column in Table 7 includes cases of snow limits stabilized at the Alps and Rocky Mountains, at the Norwegian highlands and in two cases at the Pennine and Welsh hills in Great Britain. In all these cases there were no incursions at all, even of the 17,400 ft. thickness isopleth, over the snow with less than 30 kt. winds advancing warm air over the barrier; maximum encroachments were enormously reduced even with a major thrust of warm air and high values of mean gradient wind.

/Table 7.

FOOTNOTE 1

* 18,000 feet is believed to be also about the maximum thickness observed over the inner Arctic basin in Summer at the time of gradient loosening of the pack-ice on the polar sea.

FOOTNOTE 2

* Mean maximum encroachments during a 7-day period of stabilized snow-limit for various gradient wind strengths and the absolute maximum encroachment of each thickness isopleth.

Table 7. Maximum encroachment of given thickness lines over extensive snow and ice, measured in nautical miles along the air stream. (From graphical analysis of the figures in Table 6.)

| $V_G$ (Knots) | Cases with subsidence | All Cases without subsidence (Nautical Miles) | Cases without subsidence where snow-limit was stabilized near a topographical barrier. |
|---|---|---|---|
| | | | 17,400 ft. thickness |
| 0 | 800 | 250 | 0 |
| 15 | 1,200 | 500 | 0 |
| 30 | 1,800 | 1,000 | 130 |
| 45 | 2,400 | 1,200 | (400) |
| ABS. MAX. | 2,400 | 1,500 | 900 |
| | | | 17,600 ft. thickness |
| 0 | 500 | under 200 | 0 |
| 15 | 900 | 250 | 0 |
| 30 | 1,500 | 450 | under 100 |
| 45 | 2,200 | 750 | - |
| ABS. MAX. | 2,200 | 1,100 | 500 |
| | | | 17,800 ft. thickness |
| 0 | 250 | 0 | 0 |
| 15 | 450 | under 200 | 0 |
| 30 | 900 | 200 | 0 |
| 45 | (1,400) | 350 | under 300 |
| ABS. MAX. | - | 500 | 300 |
| | | | 18,000 ft. thickness |
| 0 | 0 | 0 | 0 |
| 15 | (200) | 0 | 0 |
| 30 | (400) | 0 | 0 |
| 45 | (600) | 0 | 0 |
| ABS. MAX. | - | 0 | 0 |
| Numbers of cases used | 11 | 24 | 12 |

(c) Advance of cold air and movement of various thickness isopleths over bare ground and ocean waters beyond the snow limit. The warming of cold air over the sea has been studied by Craddock (1951) and others, and their work need not be repeated here.

Movement of the thickness isopleths for values below 17,300 ft. beyond the snow and ice margin was however briefly studied. Such movements over land are nearly always accompanied by advances of the snow limit itself. Over open ocean no such rapid changes in the distribution of frozen surface are possible. Indeed at certain times forward thrusts of cold air accompanied by high, stormy winds may cause a good deal of ice to break adrift and bring about some withdrawal of the effective ice margin.

Examination of movement of various thickness isopleths over the oceans north of 50°N yielded the following results:-

(i) Movement of the 17,400 and 17,200 ft. isopleths appears unlimited except in the southerly direction.

(ii) Extreme advances of the 17,000 ft. isopleth over the ocean may be limited to about a thousand nautical miles with extreme winds south of latitude 60°N, but are apparently unlimited farther north.

/(iii)

(iii) 16,800 ft. thickness was not observed more than 800 nautical miles from the ice margin, except north of latitude 60° where greater advances over open water occur with high winds.

(iv) 16,600 ft. and lower thickness values were not observed at all away from the ice, save with winds of over 50 kt. and in latitudes north of 60°N.

(v) The 16,200 ft. isopleth in no circumstances advanced more than 400 nautical miles from the ice.

These conclusions are based on tests in just 33 different weeks of clearly identified ice margin during six winters (1947-52).

## 6. Summary of results and concluded implications.

### (a) Results.

(1) We have found a fairly regular changeover of precipitation from unfrozen to frozen forms with 1000-500 mb. thickness about 17,300 ft. Over already established snowfields the normal thickness value for the changeover is rather higher, about 17,400 to 17,450 ft. Over seas and oceans with surface water temperature about 10°C. the average thickness value for the changeover from rain to snow is lower, about 17,150 ft. In extreme cases the changeover may come with any value from 17,100 to 17,900 ft., but in most circumstances 17,300 ± 150 ft. (5,280 ± 45m.) is the critical thickness. (The variations depend greatly on lapse rate of temperature in the lowest 1 to 2 km., but also on whether the precipitation is formed in the cloudmasses of the friction layer or at higher levels).

(2) The average 1,000 - 500 mb. thickness at the windward limit of extensive snow and ice is also found to be just 17,300 ft. with a standard deviation only about two-thirds of that characteristic for specific points normally far away from the snow/ice margin.

(3) Movement over extensive snow of the thickness isopleths for values of 17,400 ft. and higher is subject to certain restrictions. Many sorts of cooling processes are involved.

(4) When the windward limit of the snow is protected by an orographic barrier from invading warm air, advance of the higher thickness values is much more restricted. Correspondingly, in these cases, the snow-cover is more likely to persist and continue restricting the movement of the upper air pattern. Invading warm fronts are also liable to be held up and occluded at the snow-margin where this is fortified by mountains (on occasion even the hills of the British Isles) - a process which commonly yields more snow.

(5) Warming over open sea sharply restricts the movement of isopleths of the lower thickness values, when cold air thrusts out beyond the snow/ice limit. Similar forward thrusts of cold air over land usually spread the snow-cover itself to new limits.

(6) The evidence suggests that extensive snow lying itself plays some part in maintaining the cold trough in the upper troposphere which favours persistence of the snow. A snowfield of 1,500 nautical miles west to east extent - i.e. fifty degrees of longitude in 60°N or thirtyfive degrees of longitude in 45°N - would be sufficient, while it lasted, to cause a trough in the 17,400 ft. isopleth of 1000-500 mb. thickness, unless advection of warm air at an average velocity of 50 knots or more were maintained across the entire 1,500 miles or appreciable subsidence were at work. In most cases a broad trough in the thickness pattern would result. Moreover, consideration of the distribution of normal thickness values over

the northern hemisphere shows that in mid October, when the first extensive snow-deck is established over much of northeastern Canada and over eastern Asia north of the Mongolian plateau in 50-55°N, the associated cold troughs should bring the thickness down to 500 to 800 feet below the values at that time ruling over the (eastern) North Atlantic and (eastern) North Pacific oceans in the same latitude. As the winter progresses, these major cold troughs grow in longitudinal extent and in intensity and are thrust farther south, till widespread snow reaches latitude 40 to 45°N over parts of N. America and Asia in January.

These results have an obvious bearing upon the much debated question of the primacy of thermal or orographic influences in the general circulation of the atmosphere (see, for instance, Sutcliffe 1951). The investigation itself has provided instances of the effectiveness of both these factors as partial controls acting upon the circulation in the upper troposphere.

(b) Further implications. Several corollaries throw new light on known facts of meteorological experience:-

(1) For the occurrence of exceptionally prolonged and severe winters in the British Isles, the development of a continuous snow-cover right across Europe north of the Alps and Carpathians (and abnormal cooling of the North Sea surface) must be important. A snow mantle over the British Isles and western fringe of Europe alone, such as sometimes results from outbursts of maritime Arctic air from the Norwegian Sea, has far too little longitudinal extent to exercise any decisive cooling upon the atmosphere or appreciably to restrict the eastward advance of a procession of ridges and troughs in the upper westerlies. Cases of this occurred in 1944/5, 1951/2 and 1952/3, and the snow abruptly disappeared in mid winter, followed by persistently mild westerly, spells.

(2) Baur (1950) found that for Berlin winters in which the first half of December was $2\frac{1}{2}$°C. milder than normal were invariably mild winters. Development of a widespread snow-sheet across Eurasia to near Berlin by that date is likely to be a necessary, though not a sufficient, condition for persistent severe weather in western Europe.

(3) A broad continental snow-cover, as over the northern continents in winter, must favour an upper air pattern with a great warm ridge near, and just outside, the western limit of the snow and a broad cold trough over the snow. Variations of detail on this basic upper air pattern are likely to produce repeatedly situations favouring anticyclogenesis near (and over) the northwestern portion of the snow-covered region and cyclogenesis just outside the eastern or southeastern limit.

(4) An established continental snow-cover in any winter may be expected to encourage the steering of depressions along its perimeter, including southeastward-moving systems near the western limit of the snow - cf. also attempted reconstructions of ice-age phenomena by Simpson (1934) and Manley (1951).

(5) Restrictions imposed by extensive snowfields on free movement of the thickness isopleths must limit the possibilities of any radical change of depression tracks unless and until a sufficiently vigorous pulse of warm air advection removes the snow.

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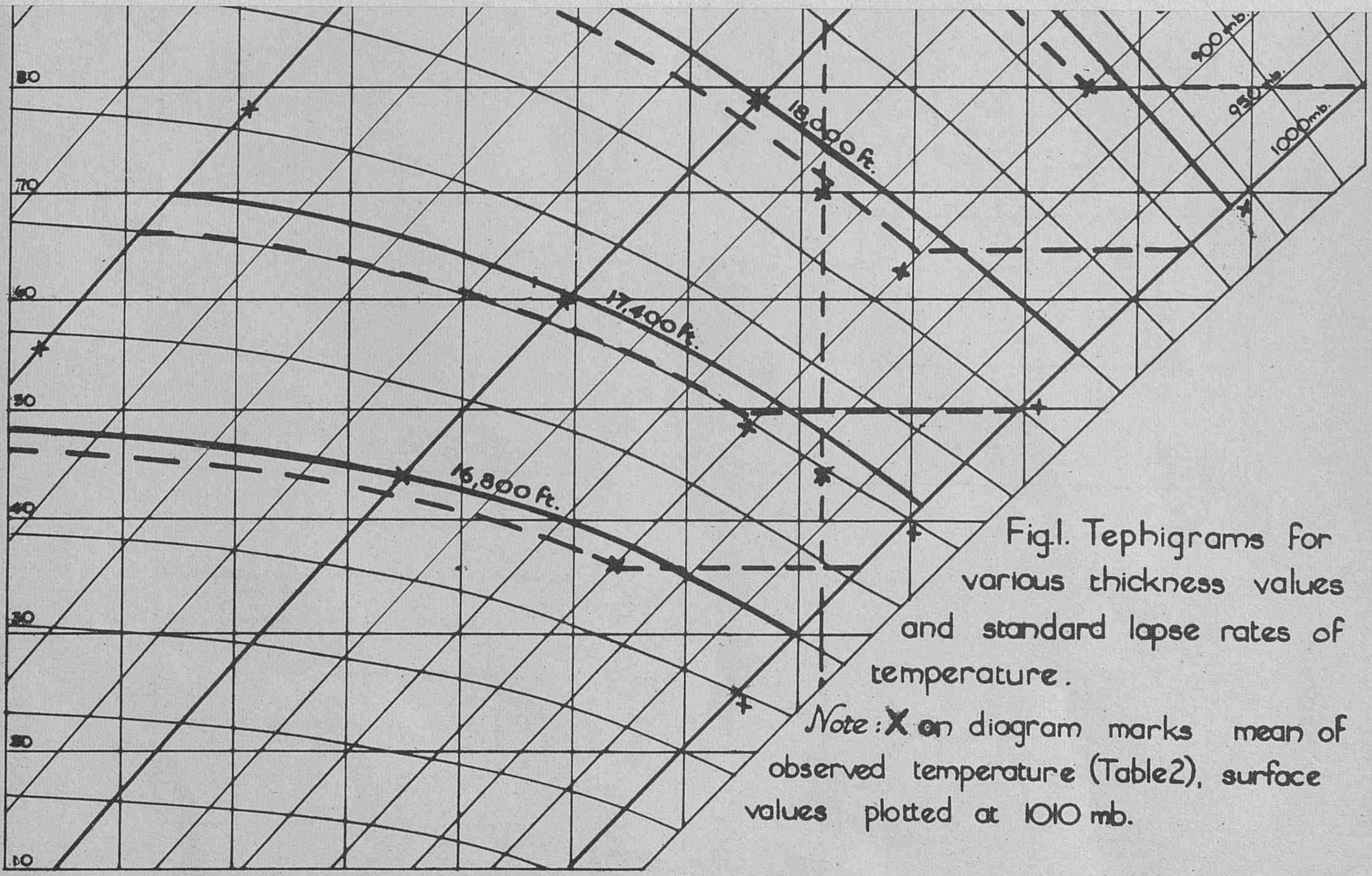

Fig.1. Tephigrams for various thickness values and standard lapse rates of temperature.

*Note*: **X** on diagram marks mean of observed temperature (Table 2), surface values plotted at 1010 mb.

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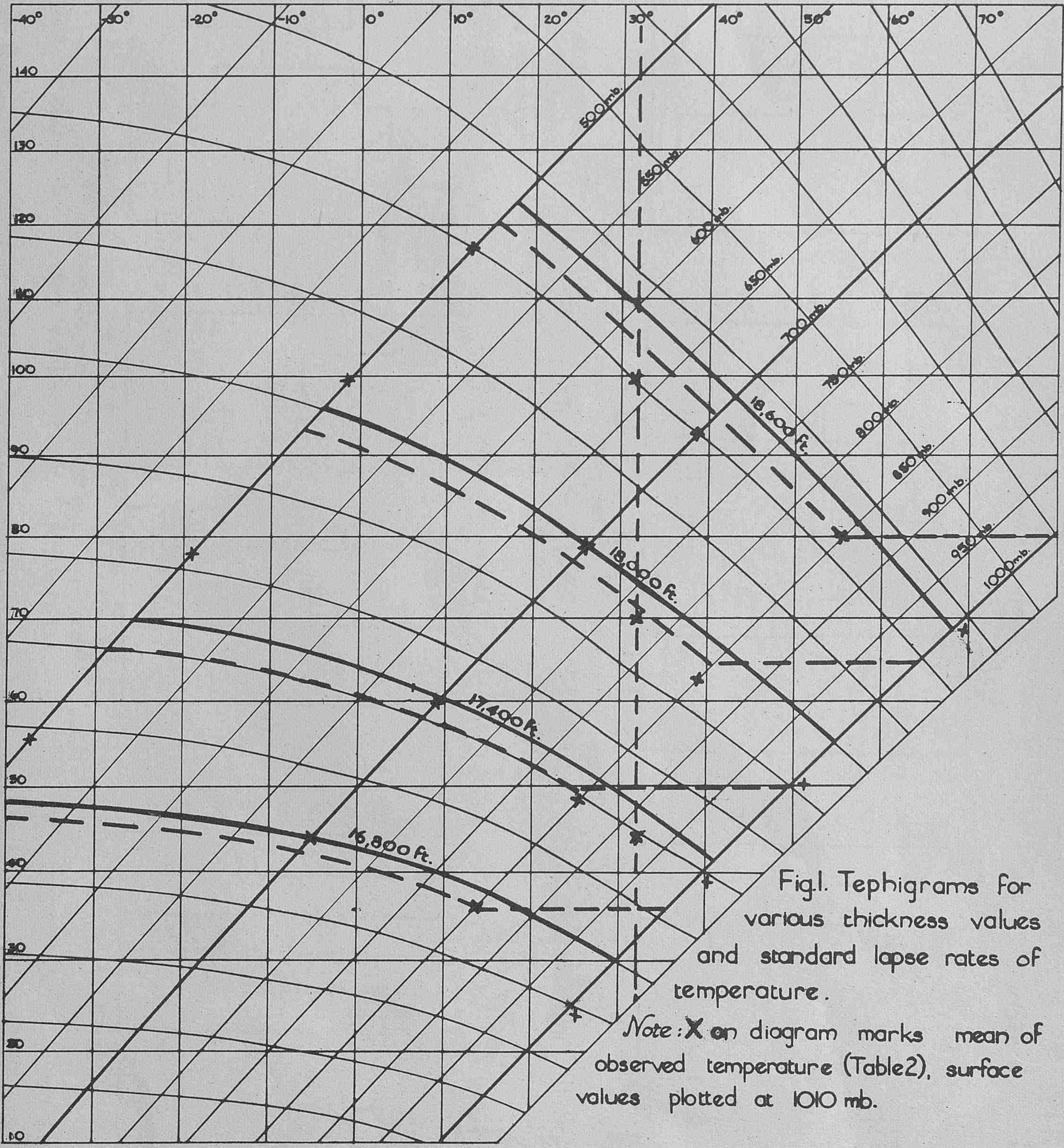

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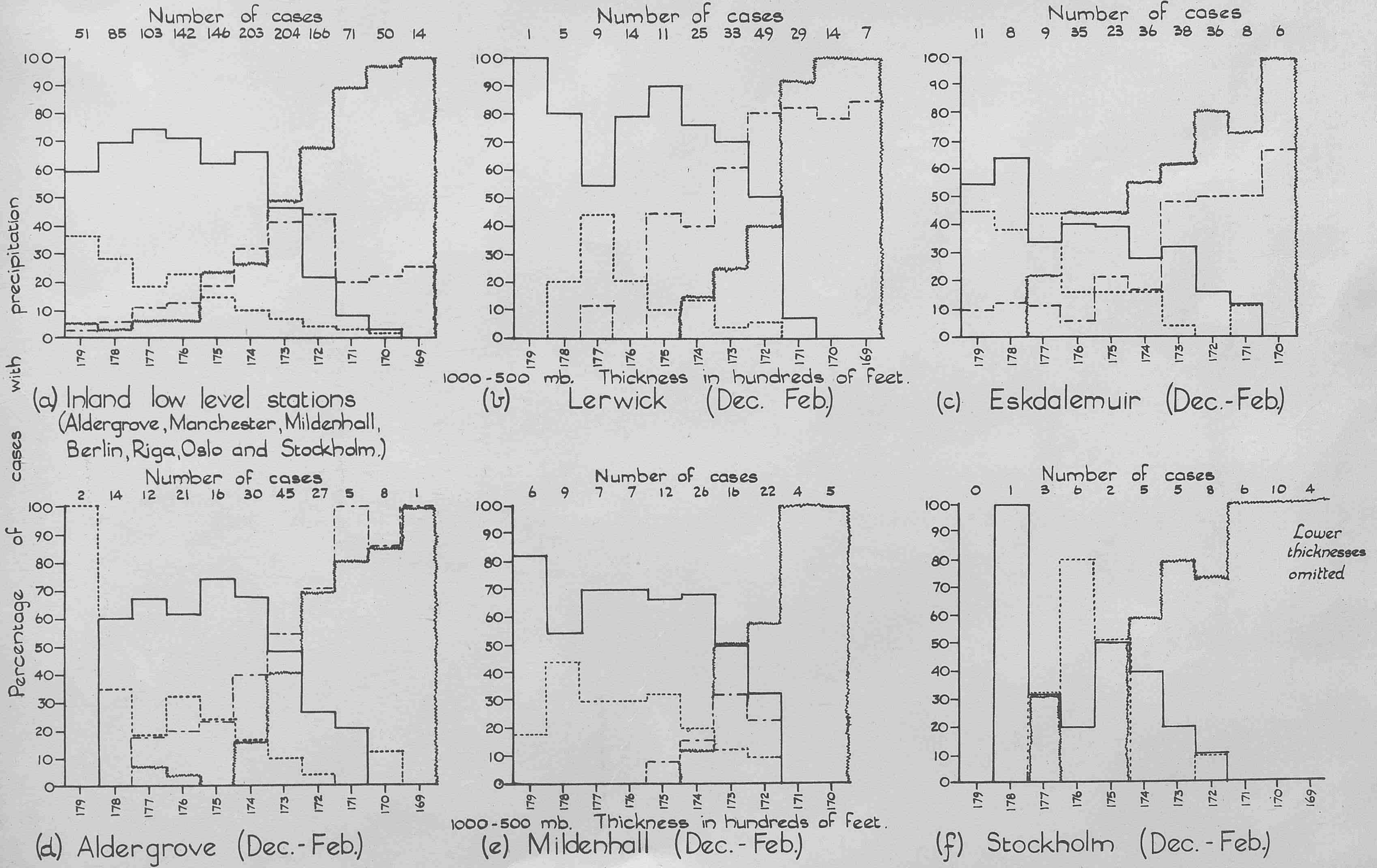

(a) Inland low level stations (Aldergrove, Manchester, Mildenhall, Berlin, Riga, Oslo and Stockholm.)

(b) Lerwick (Dec. Feb.)

(c) Eskdalemuir (Dec.-Feb.)

(d) Aldergrove (Dec.-Feb.)

(e) Mildenhall (Dec.-Feb.)

(f) Stockholm (Dec.-Feb.)

Fig. 2. Key: —— Rain ........ Drizzle ~~~~ Frozen forms (including sleet and hail) —·—·— Showers (of all forms).

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M.R.P. 836

S.C.I/81

28th October, 1953.

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AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

# Radiation and lag errors of the Meteorological Office radiosonde and the diurnal variation of upper air temperature

by

F.J. Scrase (Meteorological Office)

## Summary

Radiation errors of the M.O. radiosonde have been obtained from calculations of the rates of absorption of radiation and of heat loss by forced convection. The effect of the heating of the radiation shield on the temperature of the air reaching the thermometer element becomes very large at high levels. The errors due to direct solar radiation and to radiation reflected and emitted from below an ascending sonde have been calculated for different solar altitudes and for heights up to 30 km.

At solar altitudes between 40° and 60° the errors amount to about 1°, 2.5° and 9°C at 5, 15 and 25 km. respectively. Observed differences in readings from ascending and rapidly descending sondes indicated that the error at 25 km. was within about 7 per cent of the calculated value.

Lag coefficients have been derived from the heat transfer coefficients and the calculated value for ground level is in good agreement with laboratory measurements.

The errors of the radiosonde can account for the whole of the observed differences between day and night observations up to 15 km. Above that height, however, the differences are appreciably greater than the errors and suggest that the daily range of temperature variation over the British Isles in spring and autumn is about 1.5°C at 20 km. and about 3°C at 25 km.

1. Introduction. In M.R.P. 545 the writer described a method of calculating the radiation and lag errors of the Meteorological Office radiosonde, using the theory of heat transfer by forced convection over a flat plate. The numerical results derived in the report were based on assumed values of the absorption factors of a thermometer element and radiation shield with highly polished surfaces. In the present paper measured values, very different from the assumed values but representative of temperature units taken from stock, have been used.

The opportunity has also been taken to use more appropriate heat transfer formulae and to take into account the swinging of the radiosonde, multiple reflections in the temperature unit and the effect of radiation reflected and

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The opportunity has also been taken to use more appropriate heat transfer formulae and to take into account the swinging of the radiosonde, multiple reflections in the temperature unit and the effect of radiation reflected and emitted from the ground or from cloud. Experiments have been made to confirm some of the calculations and to compare some modifications of the radiation shield.

/2.

2. Factors affecting radiation errors. The possible effects of radiation on the temperature unit can be more readily understood by reference to the drawing of the element and shield reproduced in Fig.1A. The temperature element itself is made of polished steel/invar bimetal, 0.25 mm. thick, rolled into the shape of a hollow open-ended cylinder. In the present form of the shield, which is made of thin polished aluminium, the lower half round the element is double-walled but the inner wall is extended upwards by a rectangular duct. This arrangement was adopted after many variations had been tried out by Dymond (1947). Like all radiation shields it is necessarily a compromise, as the following considerations will show.

(a) At high solar elevations there is a possibility of solar radiation falling directly on to the element if the top opening of the shield is such as to allow a free flow of air.

(b) For the same reason radiation may reach the element after multiple reflection inside the shield.

(c) The shield itself may absorb sufficient radiation to warm appreciably the air flowing through it. (The effect of heat re-radiated from the shield to the element can be shown to be negligible).

(d) There is also the possibility of radiation reflected or emitted from the earth's surface or from clouds being strong enough to heat the temperature unit appreciably.

It is clear that the form of shield which is the more effective in preventing (a) and (b) is likely to be less effective against (c), and vice versa.

Although laboratory experiments to measure the effects of radiation under ground level conditions can be devised it is more difficult to extend such experiments to the conditions to which the radiosonde is subjected at high levels. However, the form of the element and shield adopted for the M.O. radiosonde is such as to enable the effects to be calculated by using heat transfer theory.

3. Application of theory of heat transfer. The extent to which the temperature T of a thermometer element with a heat capacity C differs from that of the air reaching it is governed by the heat transfer equation comprising the following terms:-

$$CdT/dt = Q + H + K \qquad \text{.........(1)}$$

CdT/dt being the net rate of transfer of heat and the other terms representing the rates of transfer, to or from the element, due to radiation, convection and conduction respectively. Since the supports of the radiosonde bimetal element are of very small cross-section the effect of conduction is relatively small and we can neglect K without appreciably affecting the final result. This has been confirmed by the fact that insulating the support of the thermometer element from the frame of the unit made no appreciable difference to its rate of response.

The term Q is the sum of the effects due to direct solar radiation, to radiation reflected from surfaces below the sonde and to long wave radiation emitted from below. The term H is the product of the heat transfer coefficient q and the temperature difference between the absorbing body and the air ($T-T_a$).

Equation (1) can now be re-written in terms of this temperature difference:

$$T-T_a = \frac{Q}{q} - \frac{C}{q}\,\frac{dT}{dt} \qquad \text{.........(2)}$$

The term $Q/q$ is the temperature rise due to radiation, which is thus obtainable if we know the rate of absorption of radiation and the heat transfer coefficient. The last term in equation (2) is the lag error due to thermal inertia of the body, the ratio $C/q$ being the lag coefficient.

In calculating the radiosonde errors it is necessary to distinguish between (a) the heating of the bimetal element itself and (b) the heating of the radiation shield and its effect on the air flowing through it. It is only the heating of the inner shield, with its extension beyond the outer shield, which is important, and, in what follows, any reference to the shield should be taken to mean the inner shield unless otherwise stated. The two following sections explain how the heat transfer coefficients and the rates of absorption of radiation are derived.

4. Heat transfer by forced convection. The heat transfer coefficient of a body in a moving fluid at a different temperature is given by:-

$$q = Nu\, k\, S/D \qquad (3)$$

where Nu is the Nusselt number, k the thermal conductivity of the fluid, S the area exposed to convective heat transfer and D the characteristic dimension of the body. The Nusselt number is the ratio of the rate of convective heat transfer to the rate of transfer by conduction through the fluid and the values of Nu appropriate to the form and dimensions of the body and the mass rate of fluid flow can be calculated from empirical formulae given in text-books such as that of Fishenden and Saunders (1950). The formulae given in Table I have been used in the present investigation. Re is Reynolds' number given by $v\rho D/\mu$, where v is the rate of air flow, $\rho$ the air density and $\mu$ its dynamic viscosity.

Table I

Formulae for Nusselt numbers

| Form of surface | Characteristic dimension | Re | Nu |
|---|---|---|---|
| Flat plate | Length in direction of flow. | $< 2 \times 10^4$ | $0.59\ Re^{0.5}$ |
| | | $> 2 \times 10^4$ | $0.032\ Re^{0.8}$ |
| Duct, with effective diameter d and length l | Effective diameter | $< 2.1 \times 10^3$ | $1.45\ (Re.d/l)^{0.33}$ |
| | | $> 2.1 \times 10^3$ | $0.020\ Re^{0.8}$ |
| Cylinder, perpendicular to air flow | Diameter | $10^{-1}$ to $10^3$ | $0.32 + 0.43\ Re^{0.52}$ |
| | | $10^3$ to $5 \times 10^4$ | $0.23\ Re^{0.6}$ |

The numerical constants in these formulae are dependent on Prandtl's number (given by $C_p\mu/k$, where $C_p$ is the specific heat of the fluid at constant pressure) which has the value 0.71 for air over the range of temperature and pressure in atmospheric conditions up to 30 km.

The flat plate formulae can be applied to obtain the heat transfer coefficient $q_e$ for the inner and outer surfaces of the bimetal element in the shield, the characteristic dimension D being the length in the direction of the air stream through the shield. For the coefficient $q_i$ for the inner surface of the shield the duct formulae are applicable, with the effective diameter as D. The flat plate formulae would be appropriate for the outer surface of the shield if the sonde ascended vertically. Generally, however, it swings at the end of its 40m.

suspension with an amplitude of about ± 30°, so there is an air flow perpendicular to the axis of the shield varying between 0 and 10 m/s. The average resultant air flow is about 9.5 m/s at an angle of about 45° to the axis. There is little difference between the heat transfer coefficient $q_p$, given by the flat plate formula, for an air flow parallel to the axis and the coefficient $q_c$, given by the cylinder formula, for the same rate of air flow perpendicular to the axis. The mean of $q_p$ and $q_c$ for a rate of flow of 9.5 m/s has therefore been taken as representing the coefficient $q_o$ for the outer surface of the shield.

Table II gives the dimensions used in obtaining the heat transfer coefficients.

### Table II

Dimensions used in calculating heat transfer coefficients

| Surface | Aerodynamic shape | Effective length | Width (perimeter) | Effective diam. | Surface area |
|---|---|---|---|---|---|
| | | cm. | cm. | cm. | cm.$^2$ |
| Bimetal element, inner and outer surfaces. | Flat plate | 1.6 | 3.0 | - | 9.6 |
| Shield, inner surface. | Duct | 15 | - | 3.3* | 157 |
| Shield, outer surface, flow parallel to axis. | Flat plate | 10 | 15 | - | 150 |
| Shield, outer surface, flow perpendicular to axis. | Cylinder | 8.5 | - | 4.7+ | 127 |

* Effective diameter of duct of rectangular cross-section = 4 x cross-section/perimeter

+ Effective diameter of rod of rectangular cross-section = perimeter/π

The values of the coefficients for a sonde ascending at 6 m/s and swinging through an angle of ± 30° have been calculated for various heights and are given in Table III. With the rate of ascent constant the basic variables are the density, viscosity and thermal conductivity of the air; the values of these used are included in Table III.

### Table III

Heat transfer coefficients for sonde ascending at 6 m/s and swinging ± 30°

Unit: cal. sec.$^{-1}$ °C$^{-1}$ x $10^{-3}$

| Height km. | Pressure mb. | Density $10^6\rho$ | Dyn. Visc. $10^6\mu$ | Thermal Cond. $10^6k$ | Shield | | | | | Air in shield | | Element |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | $q_p$ | $q_c$ | $q_o$ | $q_i$ | $q_s$ | m $C_p$ | B | $q_e$ |
| 30 | 12 | 18 | 147 | 49 | 15.2 | 15.8 | 15.5 | 13.0 | 25.8 | 23.8 | .420 | 1.90 |
| 25 | 25 | 39 | 146 | 49 | 21.9 | 22.3* | 22.1 | 16.3 | 36.2 | 50 | .280 | 2.77 |
| 20 | 55 | 88 | 144 | 48 | 31.4 | 34.6* | 33.0 | 20.5 | 51.7 | 106 | .175 | 3.95 |
| 15 | 115 | 184 | 145 | 48 | 46.9 | 50.8* | 48.9 | 24.1* | 71.8 | 238 | .097 | 5.9 |
| 10 | 264 | 413 | 146 | 49 | 71 | 90.7* | 81 | 45.6* | 125 | 520 | .084 | 8.9 |
| 5 | 531 | 723 | 163 | 54 | 136* | 135* | 135 | 75.5* | 208 | 946 | .077 | 12.7 |
| 0 | 1013 | 1225 | 179 | 60 | 201* | 184* | 192 | 112* | 301 | 1.530 | .071 | 16.9 |

Explanatory notes for Table III:

* The Reynolds' numbers in these cases are above the critical values given in Table I; in the other cases they are below these values.

$q_p$ refers to flow along outer surface parallel to axis.

$q_c$ refers to flow over outer surface perpendicular to axis.

$q_o$ is the mean of $q_p$ and $q_c$.

$q_i$ refers to flow over inner surface.

$q_s$ $= q_o + mC_p B$ = total transfer coefficient for shield.

$mC_p$ is the mass rate of heat flow in the air through the shield.

B $= (T_a^1 - T_a)/(T_s - T_a)$ i.e. ratio of heating of air in shield to the heating of the shield.

$q_e$ refers to flow over inner and outer surfaces of element.

As there is relatively little variation of viscosity and thermal conductivity with height, especially above 10 km., it is the air density which is almost wholly responsible for any change in the heat transfer, and therefore the radiation error, with height.

In applying the calculated heat transfer coefficients in equation (2) there is a complication arising from the fact that whereas the formulae for the Nusselt numbers for the flat plate and cylinder are based on temperature differences $(T-T_a)$ between the body and the undisturbed air upstream, the formulae for the duct refer to the mean of the differences at the inlet and outlet. Further, since the air temperature in the duct varies across the cross-section, a mean value $T_a^1$ corresponding to perfect mixing is assumed. It is this temperature which is taken to be that of the air reaching the bimetal element. Applying, then, equation (2) to the case of the shield we have:

$$C_s \frac{dT_s}{dt} = Q_s - q_o (T_s - T_a) - q_i \left[T_s - (T_a + T_a^1)/2\right] \qquad \text{.........(4)}$$

The last term, representing the rate of heat loss by the inner surface, must be equal to the rate of gain of heat by the air passing through the shield, which is $mC_p (T_a^1 - T_a)$, where m is the mass rate of air flow, i.e. $\pi D^2 \rho v/4$, D being the effective diameter of the duct. It follows that:

$$\frac{T_a^1 - T_a}{T_s - T_a} = \frac{2q_i}{2mC_p + q_i} = B \text{ (say)} \qquad \text{.........(5)}$$

thus giving the ratio of the heating of the air to the heating of the shield. It also follows that:-

$$q_i \left[T_s - (T_a + T_a^1)/2\right] = mC_p\, B\, (T_s - T_a) \qquad \text{.........(6)}$$

We can therefore now express the heat balance of the shield by:

$$C_s \frac{dT_s}{dt} = Q_s - q_s (T_s - T_a) \qquad \text{.........(7)}$$

where $q_s$, the total heat transfer coefficient, is given by:

$$q_s = q_o + mC_p B$$

The calculated values of $q_o$, $mC_p$, B and $q_s$ are included in Table III. An experimental determination of the value of B was attempted by measuring the temperature of air passing through a heated shield in a wind tunnel. The shield was maintained about 5°C warmer than the air stream, the speed of which was 6 m/s. Measurements were made with the axis of the shield not only in the direction of the air flow but also displaced 30° in horizontal and vertical planes. Temperatures were measured by thermocouples but as the experiments could only readily be made at ground level pressure the temperature differences between the air entering and leaving the shield were very small and there was a large scatter in the individual readings. The mean, however, of 24 measurements gave the value of B as 0.08 ± 0.08 P.E. which is not inconsistent with the calculated value of 0.071.

5. Rate of absorption of radiation. The radiation term Q in the heat balance equation depends, of course, on the intensity of the incident radiation I, on the absorption factor A of the body and on the orthogonal projection of the area exposed to the beam. The computation of Q is complicated by (a) the swinging of the sonde, (b) shading of the shield or element by other parts of the sonde, and (c) multiple reflections of radiation entering the open ends of the shield.

For the absorption of solar radiation of intensity I by the element or the shield we can write:

$$Q_h = AI \cos h . f(S_d, h, A_s) \qquad \text{.........(8)}$$

where h is the solar altitude, $S_d$ the effective vertical area exposed, including that receiving radiation reflected internally, and $A_s$ is the absorption factor of the shield.

The numerical values of f $(S_d, h, A_s)$ are best obtained by geometrical measurement.

The rate of absorption of radiation reflected from the ground or from cloud is given by:

$$Q_m = AI_r \int_{\alpha_1}^{\alpha_2}\int_{\phi_1}^{\phi_2} \cos^2\phi . d\phi . S_m d\alpha \qquad \text{.........(9)}$$

where $S_m$ is the effective vertical area of the shield or element exposed, which may vary with the azimuth and elevation angles, $\alpha$ and $\phi$, of the effective portions of the radiating surfaces as "seen" by the sonde. $I_r$ is the intensity of the reflected radiation and is given by:

$$I_r = RI \sin h \qquad \text{.........(10)}$$

where R is the albedo of the reflecting surface and I is the intensity of solar radiation at the surface.

The rate of absorption of radiation $Q_1$ emitted from below is also given by (9) except that the intensity of the long wave radiation $I_1$ and the appropriate absorption factor $A_1$ must be used. The intensity $I_1$ is given by

$$I_1 = \sigma (T_o^4 - T^4) / \pi \qquad \text{.........(11)}$$

where $\sigma$ is the Stefan-Boltzmann coefficient (1.37 x $10^{-12}$ cal.cm.$^{-2}$sec.$^{-1}$°C$^{-4}$) and $T_o$ and T are the temperatures of the emitting surface (cloud or ground) and the sonde. The element and shield themselves emit a negligible amount of radiation to their immediate surroundings.

The angular limits $\alpha_1$, $\alpha_2$, $\phi_1$, $\phi_2$, in (9) are governed by the extent of the shading of the element and the shield by other parts of the sonde. For evaluating $Q_m$ and $Q_l$ the integrations with respect to $d\phi$ have been done in 10° steps, and the effect of swinging of the sonde has been taken into account. The intensity of solar radiation is taken to be 0.032 cal. cm.$^{-2}$ sec.$^{-1}$ at the top of the atmosphere but in the final results it is adjusted for air mass appropriate to the solar altitude and the height of the sonde by means of a graph provided by Vaisala (1948). In evaluating $I_r$ a value of 0.35 for R has been taken as representing average conditions of the earth's surface and cloud cover for the British Isles. The long wave radiation $I_l$ has been calculated for a surface temperature $T_o$ of 288°K and a sonde temperature T of 220°K; these conditions cover most soundings in the British Isles except at heights below 10 km.

The absorption factors of the temperature element and the shield were measured by the National Physical Laboratory at various wave lengths between 0.3 and 10 microns for both normal and 45° incidence, a random sample of each material being taken from stock. From these measurements and from the spectral distribution of solar energy the absorption factors were found to be 0.33 for the aluminium shield and 0.75 and 0.60 for the two sides of the bimetal element. A mean of 0.67 has been used for the element. The values for 45° incidence did not differ from those at normal incidence by more than 0.02. The N.P.L. values for a wave length of 10 microns, viz. 0.03 for the shield and the mean of 0.18 for the element have been used for the absorption factors for the long waves. It must be borne in mind that there may be appreciable differences between individual samples, depending on the state of their surfaces. For a white card shield used in some experimental soundings to be described later, factors, based on published data, of 0.25 for solar radiation and 0.95 for long waves have been used.

Calculated rates of absorption of radiation by the shield and by the element inside it for different solar altitudes are given in Table IV.

Table IV

Rate of absorption of radiation in cal.sec.$^{-1}$ x $10^{-3}$

| Solar altitude | h = | 10° | 30° | 50° | 70° | 90° |
|---|---|---|---|---|---|---|
| Shield | | | | | | |
| Direct solar: | $Q_h$ | 543 | 597 | 543 | 422 | 415 |
| Reflected from below: | $Q_m$ | 32 | 93 | 141 | 174 | 185 |
| Emitted from below: | $Q_l$ | 8 | 8 | 8 | 8 | 8 |
| Total: | $Q_s$ | 583 | 698 | 692 | 604 | 608 |
| Element | | | | | | |
| Direct solar: | $Q_h$ | 0.9 | 4.1 | 7.8 | 9.4 | 12.2 |
| Reflected from below: | $Q_m$ | 0.7 | 2.0 | 3.0 | 3.7 | 4.0 |
| Emitted from below: | $Q_l$ | 0.5 | 0.5 | 0.5 | 0.5 | 0.5 |
| Total: | $Q_e$ | 2.1 | 6.6 | 11.3 | 13.6 | 16.7 |

Note: The figures for the absorption of reflected radiation are for an albedo 0.35 and those for long wave radiation refer to a surface at 288°K with the sonde at 220°K.

The values for direct radiation take account of the swinging of the sonde through an angle of 30°. This swinging makes no appreciable difference to the amount of radiation reaching the sonde from below. The values are plotted in Fig.2 for comparison with values relating to a sonde remaining vertical. The curves are similar in shape to those given by Vaisala (1941) for his design of sonde.

The effect of radiation emitted from below is quite negligible in the case of the shield, but the contribution by the reflected radiation to the total absorption increases from near zero at low solar altitudes to 30 per cent at 90°. In the case of the element the very small amount of radiation absorbed at low solar altitudes is roughly half due to the direct radiation entering the upper opening of the shield and half to radiation reflected or emitted from below. But at high solar altitudes the direct radiation contributes about three-quarters and reflected radiation one quarter to the total, the effect of the emitted radiation then being less than five per cent. It must be remembered that at low solar altitudes the intensity of the sun's radiation is reduced appreciably below the value 0.032 cal. $\text{cm.}^{-2}\text{sec.}^{-1}$ assumed in Table IV. At solar altitudes above 20° and heights about 10 km. the reduction is less than 10 per cent.

6. Calculated radiation errors. The values of $q_s$, $q_e$, B, $Q_s$ and $Q_e$ given in Tables III and IV enable us to obtain the radiation terms $Q_s/q_s$ and $Q_e/q_e$ in the equations:

$$T_s - T_a = \frac{Q_s}{q_s} - \frac{C_s}{q_s}\frac{dT_s}{dt} \quad \text{and} \quad T_e - T_a{}^1 = \frac{Q_e}{q_e} - \frac{C_e}{q_e}\frac{dT_e}{dt} \qquad \text{.........(12)}$$

But whereas $Q_s/q_s$ gives the radiational heating of the shield above the free air temperature, $Q_e/q_e$ only gives the heating of the element above the temperature $T_a{}^1$ of the air in the shield. The definition of B in equation (5), however, enables us to eliminate $T_a{}^1$ and obtain the heating of the element above the free air temperature, thus:

$$T_e - T_a = \frac{Q_e}{q_e} + \frac{BQ_s}{q_s} - \frac{C_e}{q_e}\cdot\frac{dT_e}{dt} - \frac{BC_s}{q_s}\cdot\frac{dT_s}{dt} \qquad \text{.........(13)}$$

The terms $(Q_e/q_e + BQ_s/q_s)$ together represent the total radiation error of the bimetal element in the shield. The calculated values of this error for a selection of heights and solar altitudes are given in Table V, which also includes values for the heating of the shield and the air in it.

/Table V

## Table V

### Calculated radiation errors of the M.O. radiosonde

(Combining the effects of direct solar radiation, reflected radiation from earth and/or cloud with average albedo 0.35, and long wave radiation emitted from the earth and/or cloud).

| Solar Altitudes | | 10° | 30° | 50° | 70° | 90° |
|---|---|---|---|---|---|---|
| | Height km. | | | °C | | |
| (a) Heating of shield $Q_s/q_s$ | 30 | 22.6 | 27.1 | 26.9 | 23.4 | 23.6 |
| | 20 | 11.3 | 13.5 | 13.4 | 11.7 | 11.8 |
| | 10 | 4.7 | 5.6 | 5.6 | 4.8 | 4.9 |
| | 0 | 1.9 | 2.3 | 2.3 | 2.0 | 2.0 |
| (b) Heating of air in shield $BQ_s/q_s$ | 30 | 9.6 | 11.4 | 11.3 | 9.9 | 10.0 |
| | 20 | 2.0 | 2.4 | 2.3 | 2.0 | 2.1 |
| | 10 | 0.4 | 0.5 | 0.5 | 0.4 | 0.4 |
| | 0 | 0.1 | 0.2 | 0.2 | 0.1 | 0.1 |
| (c) Error of element with respect to air in shield $Q_e/q_e$ | 30 | 1.1 | 3.5 | 6.0 | 7.2 | 8.8 |
| | 20 | 0.5 | 1.7 | 2.9 | 3.4 | 4.2 |
| | 10 | 0.2 | 0.7 | 1.3 | 1.5 | 1.9 |
| | 0 | 0.1 | 0.4 | 0.7 | 0.8 | 1.0 |
| (d) Error of element with respect to free air $Q_e/q_e + BQ_s/q_s$ | 30 | 10.6 | 14.9 | 17.3 | 17.1 | 18.8 |
| | 25 | 5.2 | 7.7 | 9.3 | 9.5 | 10.7 |
| | 20 | 2.4 | 4.0 | 5.1 | 5.3 | 6.2 |
| | 15 | 1.1 | 1.9 | 2.7 | 3.0 | 3.5 |
| | 10 | 0.5 | 1.1 | 1.7 | 1.8 | 2.2 |
| | 5 | 0.3 | 0.7 | 1.1 | 1.2 | 1.4 |
| | 0 | 0.1 | 0.4 | 0.7 | 0.7 | 0.9 |

NOTE: In (a), (b) and (c) the intensity of solar radiation is taken to be 0.032 $\mathrm{cal.cm.}^{-2}\mathrm{sec.}^{-1}$. In (d) account is taken of the effect of air mass on the intensity.

It will be seen that for solar altitudes about 50° the heating of the air in the shield is responsible for about 20 per cent of the radiation error near the ground, 30 per cent at 10 km. and 65 per cent at 30 km. The contributions are higher still at lower solar altitudes since the angle of incidence of solar radiation on the shield is nearer to normal incidence. The remainder of the error is, of course, due to radiation falling on the element itself, either directly or after reflection by the inner surface of the shield.

7. Lag errors. The lag in temperature of the shield and the element is represented by the last terms in equations (12), $C/q$ being the lag coefficient which is usually written as $\lambda$. Since $C$ is readily obtained from the mass of the body and its specific heat we can obtain $\lambda$ from the calculated value of $q$. The lag coefficient can also be measured in the laboratory (where radiational exchange can be made negligible) by observing the time required for an initial temperature difference between the thermometer element (or shield) and the air to be reduced in the ratio $1/e$. In such conditions

$$T - T_a = (T_o - T_a)\exp - \frac{t}{\lambda} \qquad (14)$$

where $T = T_o$ at $t = 0$. This measurement of $\lambda$ therefore affords a check on the calculated heat transfer coefficient.

For the bimetal element without its shield the measured value of $\lambda_e$ in an air flow of 6 m/s at ground level pressure was 7.3s. Under the same conditions the calculated value of $q_e$ of $16.9 \times 10^{-3}$ (from Table III), with C = 0.116, gives $\lambda_e = 6.9$s. The agreement is reasonable.

When the thermometer element is in its shield, however, its lag behind the free air temperature will be affected by the lag of the shield itself and by the extent to which the shield heats the air reaching the element. The total lag error is, in fact, given by the last two terms of equation (13), but it is not readily evaluated since $dT_s/dt$ is not normally measured. The lag of the element in the shield can, however, be measured in the laboratory. From equation (14) and equations (12) for the shield, together with the definition of B, it can be shown that for a sudden temperature difference in the absence of radiational exchange:

$$\frac{dT_e}{dt} = \frac{1}{\lambda_e} \left[ T_e - T_a - B(T_o - T_a) \exp \frac{-t}{\lambda_s} \right] \qquad \text{.......(15)}$$

and that:

$$T_e - T_a = (T_o - T_a) \left( \exp \frac{-t}{\lambda_e} \right) \left[ 1 + \frac{B\lambda_s}{\lambda_s - \lambda_e} \left\{ \exp\left( \frac{t}{\lambda_e} - \frac{t}{\lambda_s} \right) - 1 \right\} \right] \text{(16)}$$

Now when $\lambda_s$ is not very different from $\lambda_e$, which is the case with the radiosonde unit, and when the measurements are made with t not greatly exceeding $\lambda_e$, this simplifies to:

$$T_e - T_a = (T_o - T_a) \exp - \frac{t(1 - B)}{\lambda_e} \qquad \text{.......(17)}$$

In other words the lag coefficient of the shielded element $\lambda_{es}$ is $1/(1-B)$ times that of the unshielded element and the total lag error can now be written:

$$\lambda_{es} \frac{dT_e}{dt} \qquad \text{or} \qquad \frac{\lambda_e}{1-B} \cdot \frac{dT_e}{dt}$$

The laboratory measurement of $\lambda_{es}$ under the same conditions as for $\lambda_e$ gave a value of 7.6s. The calculated value given by $C_e/q_e(1-B)$ is 7.4s.

It is difficult to measure the lag coefficient in laboratory conditions corresponding with the normal rate of ascent at very low air densities, but it is reasonable to expect that the method of calculation which has been found to give satisfactory agreement at ground level should also be satisfactory for high level conditions. Table VI gives the lag coefficients calculated for various heights and the normal rate of ascent of 6 m/s. The values of the heat capacities used were 0.116 at 288°K and 0.106 at 220°K for the element and 2.42 and 2.16, at these temperatures, for the shield.

/Table VI

## Table VI

### Lag coefficients and errors (rate of ascent 6 m/s)

| | Coefficients, in seconds | | | Typical errors (°C) | |
|---|---|---|---|---|---|
| Height | $\lambda_e$ | $\lambda_s$ | $\lambda_{es}$ | night | day |
| 30 | 56 | 84 | 97 | -0.6 | -1.6 |
| 25 | 38 | 60 | 54 | -0.3 | -0.7 |
| 20 | 27 | 42 | 33 | -0.2 | -0.3 |
| 15 | 18 | 30 | 20 | | |
| 10 | 11.9 | 17.3 | 12.9 | +0.5 | +0.5 |
| 5 | 8.7 | 11.0 | 9.5 | +0.3 | +0.3 |
| 0 | 6.9 | 8.0 | 7.4 | +0.3 | +0.3 |

NOTE: The lag errors are for a temperature gradient of -6°C/km. below 15 km. and +1°C/km. above 15 km. The daytime errors take account of the rate of change of temperature of the element by radiational heating for solar altitude 40°.

The table also includes the lag errors for typical rates of change of indicated temperature in night and day soundings. In the troposphere the errors are the same sign as, and about one third the magnitude of, the radiation errors. In the stratosphere they are very small in comparison with the radiation errors and are of opposite sign.

8. Observed errors at high levels. Some comparisons of temperatures indicated at the same height by a radiosonde ascending at the normal rate and descending at a much higher rate have afforded some confirmation of the existence of the large radiation errors at high levels. Such comparisons were made in five soundings at Downham Market (52°36'N., 0°24'E.) in September, 1950 when the balloons burst at heights near 30 km. and the parachutes failed to open immediately. The sondes were tracked by radar. Fig.3 is a graph of the readings of temperature and height obtained during one of the soundings and shows the sudden fall in the temperature readings during the rapid descent.

In the descent the heating, by the shield, of the air reaching the element is negligible since the latter is near the bottom opening of the shield. The temperature $T_d$ indicated by the element during the descent is therefore given by:

$$T_d - T_a = \frac{Q_e}{q_d} - \frac{C_e}{q_d} \cdot \frac{dT_d}{dt} \qquad \text{.......... (18)}$$

where $q_d$ is now the much larger heat transfer coefficient of the element for the much higher rate of air flow V.

Combining equations (13) and (18) then gives the difference between the ascent and descent temperatures:

$$T_e - T_d = \frac{Q_e}{q_e} + \frac{BQ_s}{q_s} - \lambda_{es}\frac{dT_e}{dt} - \frac{Q_e}{q_d} + \lambda_d \cdot \frac{dT_d}{dt} \qquad \text{.......... (19)}$$

This can be evaluated from the calculated values of the radiation errors and the lag coefficients and the observed rates of change of indicated temperatures. The results of the calculations and comparisons are given in Table VII.

/Table VII

## Table VII

### Observed and calculated errors (solar altitude 40°)

| Date 1950 Sept. | Height km. | Rate of descent m/s | Calculated errors in °C | | | | | | Ascent - descent $(T_e - T_d)$ | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | Ascent | | | Descent | | | | | |
| | | | Radiation | Lag | $(T_e - T_a)$ | Radiation | Lag | $(T_d - T_a)$ | Calc. | Obs. | O-C |
| 12 | 24.0 | 41 | 7.9 | 0.2 | 7.7 | 1.2 | -0.7 | 1.9 | 5.8 | 6.1 | +0.3 |
| " | 21.8 | 34 | 6.1 | 0.3 | 5.8 | 1.1 | -0.2 | 1.3 | 4.5 | 5.2 | +0.7 |
| 13 | 28.2 | 57 | 13.4 | 2.8 | 10.6 | 1.4 | -1.3 | 2.7 | 7.9 | 8.8 | +0.9 |
| " | 25.8 | 47 | 9.8 | 1.3 | 8.5 | 1.3 | -0.6 | 1.9 | 6.6 | 6.0 | -0.6 |
| 19 | 28.8 | 61 | 14.5 | 2.0 | 12.5 | 1.4 | -2.9 | 4.3 | 8.2 | 10.3 | +2.1 |
| " | 25.2 | 44 | 9.1 | 1.3 | 7.8 | 1.3 | -1.1 | 2.4 | 5.4 | 5.7 | +0.3 |
| 20 | 27.8 | 55 | 12.7 | 1.3 | 11.4 | 1.4 | -2.1 | 3.5 | 7.9 | 9.4 | +1.5 |
| " | 24.3 | 42 | 8.2 | 0.5 | 7.7 | 1.2 | -0.7 | 1.9 | 5.8 | 4.5 | -1.3 |
| 21 | 27.1 | 52 | 11.6 | 1.1 | 10.5 | 1.3 | -1.4 | 2.7 | 7.8 | 8.9 | +1.1 |
| " | 23.3 | 38 | 7.2 | 0.4 | 6.8 | 1.2 | -0.2 | 1.4 | 5.4 | 6.0 | +0.6 |
| Mean | 25.6 | 47 | 10.0 | 1.1 | 8.9 | 1.3 | -1.1 | 2.4 | 6.5 | 7.1 | +0.6 |

It should be remembered that the observations for each height are single pairs of readings and considerable scatter in the results must be expected. The calculation of the lag error in the descent is probably the most doubtful factor in the computations, since the rapid rate of change of temperature is difficult to measure accurately. However, the mean difference of 0.6°C between the observed and calculated temperature differences can be considered reasonably good and it might well be accounted for by the albedo being different from the value 0.35 assumed in the calculations. The soundings were made with the sky about half clouded but the surface exposed to the sondes through the breaks in the cloud was part grassland and part sea.

9. Comparisons with modified shields. In order to see whether radiation errors could be appreciably reduced by modifying the design of the shield a series of comparisons of two sondes, one standard and one modified, on the same balloons was carried out at Downham Market in September, 1950. The modifications used are shown in Fig. 1. The series also included sondes without any shield, i.e. with the element fully exposed except for partial shading by the sonde. Five or six soundings to high levels were made with each modification. Calculated radiation errors for the standard and for the modified sondes at solar altitudes 40°, are plotted in Fig. 4. The results of the comparisons are given in Fig. 5 in which the points on the graphs represent the means of about 5 observations of the differences (except at the highest levels where fewer were obtained). Probable errors of the mean differences are also indicated. The calculated differences are derived from Fig. 4 and do not include the lag errors, which, however, would not affect the observed differences by more than 0.3°C. The agreement between the observations and the calculations is generally within 1°C. The differences that the modifications make, especially below 20 km., are relatively small. Dymond (1947) found much the same result in some earlier experiments with modified shields. It emphasises the fact, already mentioned, that the design has to be a compromise between conflicting requirements. Thus, reducing the length of the shield, as in modification B, decreases the heating effect of the shield but exposes the element more to direct radiation. Further decrease of the radiation falling on the shield, by extending the outer shield as in modification C, appears to be nullified by decrease in the heat convected from the outer surface of the inner shield. Moreover, neither B nor C can make any difference to the amount of radiation absorbed

inside the shield. The effect of the heating of the standard pattern shield increases so rapidly at high levels that at 30 km. the radiation error is as large as when the element is fully exposed. The only modification that showed any appreciable improvement - and this only at levels above 20 km. - was D, in which the inner shield was made of white card instead of aluminium. This is probably due to the card having a lower absorption factor for solar radiation and to its being a diffuse reflector and so reflecting less radiation on to the element.

10. The diurnal variation of upper air temperature. If the radiation and lag errors are correctly evaluated the magnitude of the diurnal range of air temperature variation, and its change with height, should be obtainable from the differences between day and night radiosonde observations. In Fig. 6 the curve gives the combined radiation and lag errors for a solar altitude of 40°. In the same graph are plotted four series of mean differences between night and day observations, the latter being made at about this solar altitude. With the exception of those derived from three pairs of high altitude soundings at Downham Market on 15th, 25th and 29th September, 1950, the means are for routine soundings for the six months February to April and August to October combined. Those published by Kay (1951) represent about 1800 observations in 1947 and 1948 at five stations in the British Isles. Piegza's (1947) means are derived from about 700 observations in 1943-6 at Fazakerley (53°28'N., 2°55'W.). The means for the routine soundings at Downham Market represent about 1,400 observations in the period 1943-51. It is clear that up to about 15 km. radiation errors could account for the whole of the day and night differences. Above that height the comparisons indicate that the differences are too big to be accounted for wholly by the errors, if these are correctly assessed, and that there is a diurnal range of about 1.5°C at 20 km. (55 mb.) increasing to about 3°C at 25 km. (25 mb.). We cannot, of course, regard the day and night differences above 22 km. as representative of average conditions since they were based on only three comparisons and although each was made on the same day, at 10h. and 22h., they may well be affected by day to day temperature changes. On the other hand much the same differences above 22 km. were obtained in the means of 15 day and 15 night soundings in September and October which were not made on the same days.

The diurnal ranges at 20 km. and 25 km. indicated by these results are ten times as large as those recently calculated by F.S. Johnson (1953) from the heating due to absorption of solar energy by atmospheric ozone. He used direct measurements of ozone concentration and ultra-violet spectral intensity of sunlight during a rocket flight over New Mexico in June, 1949 and, assuming a constant rate of energy loss, found the diurnal temperature change to have a maximum value of 5.3°C at 48 km. falling to 1°C at 31 km. and 70 km.

Goldie (1951) has suggested the possibility of the diurnal temperature change being due to dynamical causes, in which case the diurnal range should vary approximately inversely with the pressure (or density) and should, therefore, increase exponentially with height. An adiabatic heating of 1.5°C at 20 km. would require a pressure increase of 1.3 mb. above that level and this same increase would also account for the heating of 3°C at 25 km. An inflow of 1 or 2 mb. of air above these levels, with a corresponding outflow at low levels, is not unreasonable. Clearly, however, further information at high levels is required before any safe conclusions about diurnal variation can be drawn.

11. Conclusion. Little advantage is to be gained by introducing any further modifications to the radiation shield as they cannot reduce the radiation errors to negligible proportions. A better course is to use a temperature element for which the ratio of the rates of absorption of radiation and heat convection is so small that the shield becomes unnecessary. The factors favouring this are (a) low absorption factor, (b) small effective surface area exposed to radiation, and (c) large characteristic dimension for the Nusselt number.

In general, (b) and (c) are mutually antagonistic, but a thin plate which would favour (c) would also favour (b) if its edge could be kept facing the source of radiation. This, of course, is impracticable in a radiosonde and probably the best compromise in a thin wire element perpendicular to the air flow. Calculations show that the maximum radiation error of such an element of diameter $10^{-3}$cm. at 30 km. would be only 0.35°C even if its absorption factor were unity. An element of this type, using wire of $1.35 \times 10^{-3}$cm. diam., is to be used in the form of a coiled coil in the M.O. radar sonde and it is hoped that the radiation errors will be small enough to allow the magnitude of the diurnal variation of temperature to be determined accurately.

In the meantime, Table V (d) and the lag coefficient $\lambda_{es}$ given in Table VI can be used for applying corrections to daytime observations obtained with the M.O. radiosonde.

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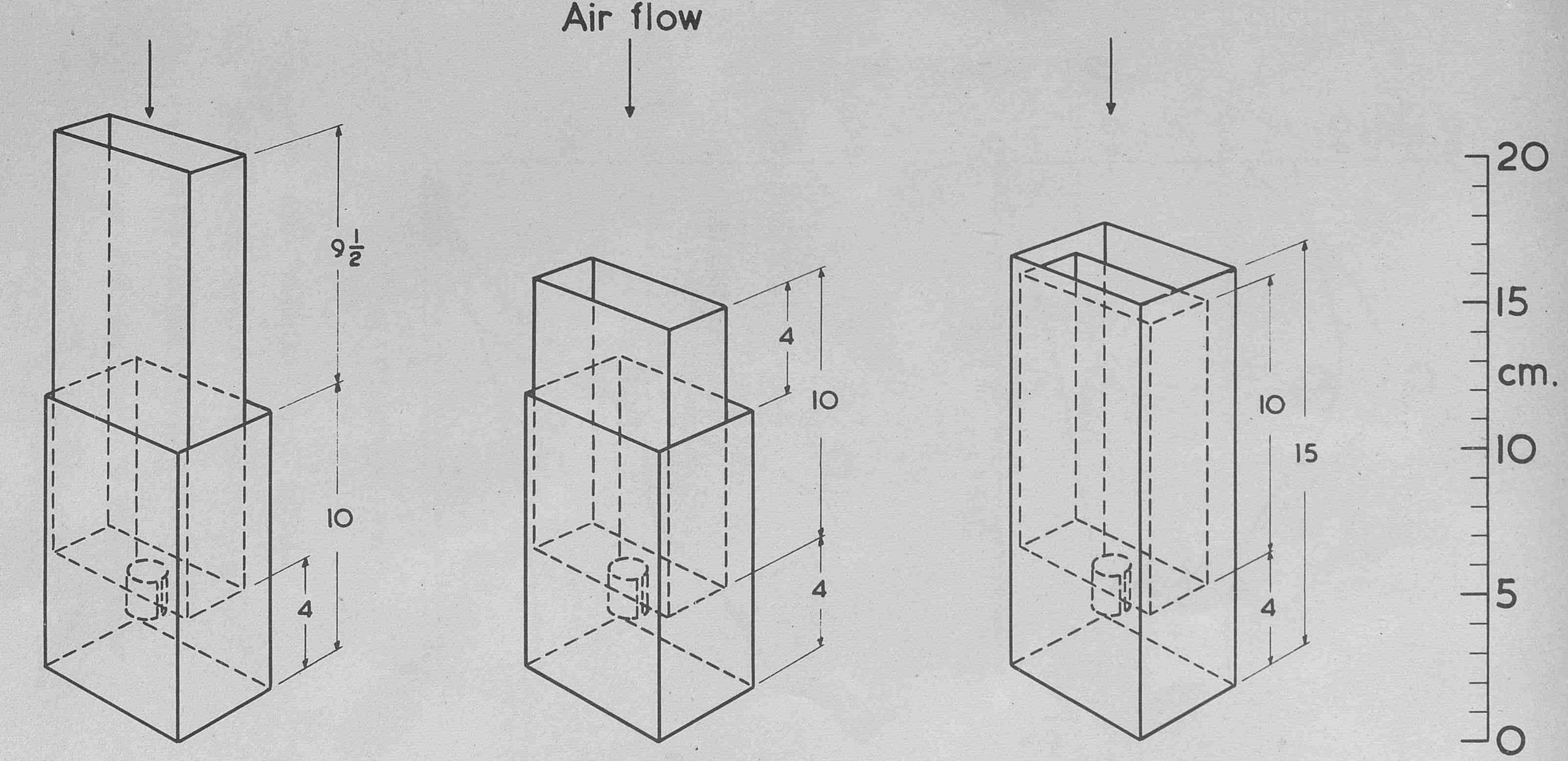

Air flow
9 1/2
10
4
4
10
4
10
15
4
20
15
cm.
10
5
0

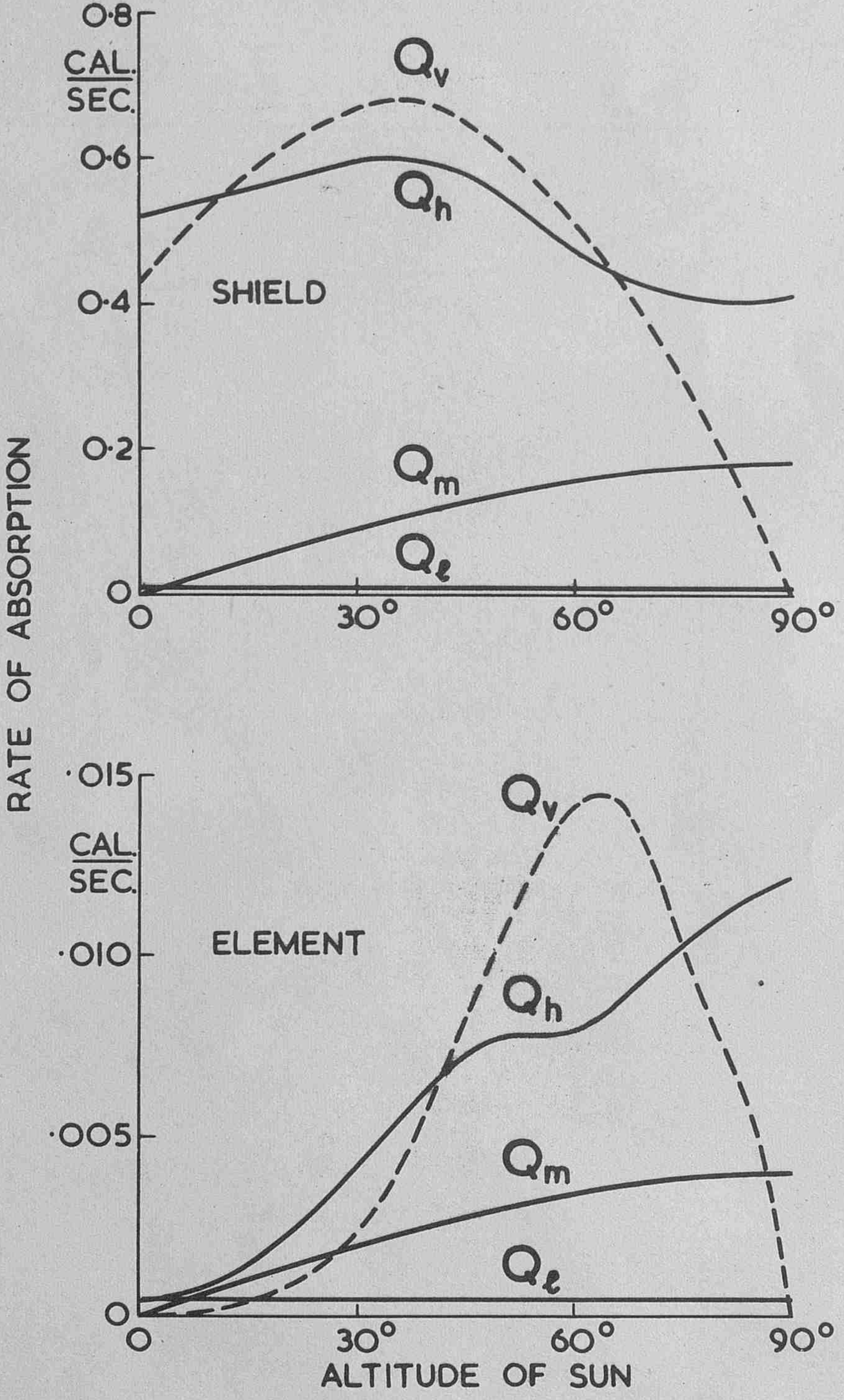
A. Current type

B. Former type

C. Inner shield aluminium

D. Inner shield white card

Fig. 1 Radio sonde radiation shields



$Q_v$ direct solar radiation; sonde vertical

$Q_h$ direct solar radiation; sonde swinging ± 30°

$Q_m$ reflected solar radiation; albedo 0·35

$Q_\ell$ long wave radiation; earth at 288°K, sonde at 220°K

(Effect of swinging on $Q_m$ and $Q_\ell$ is negligible)

Fig. 2 Radiation absorbed by element and shield

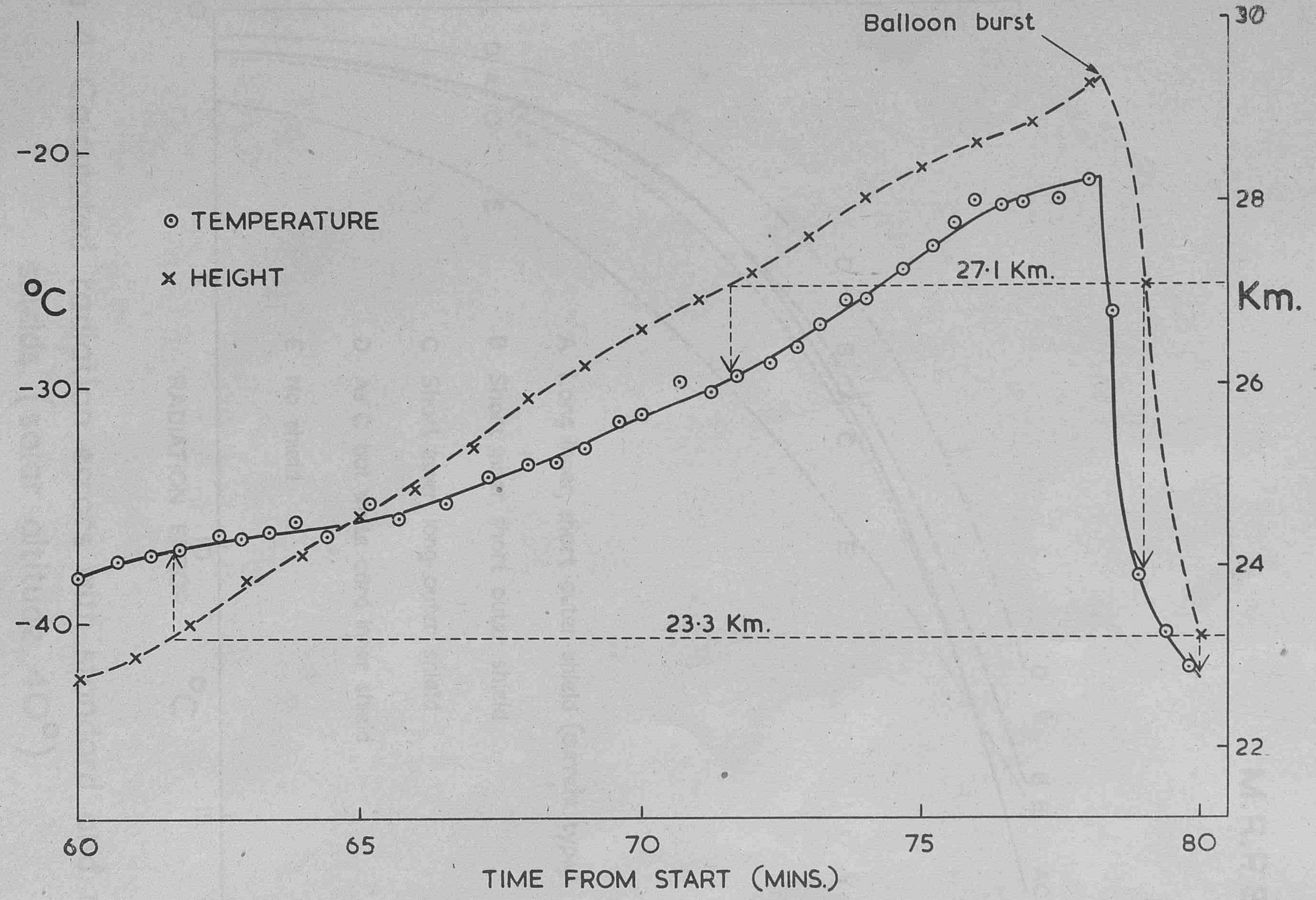


Fig. 3 Temperatures and heights observed in ascent and descent at Downham Market 10h. 21 Sept. 1950

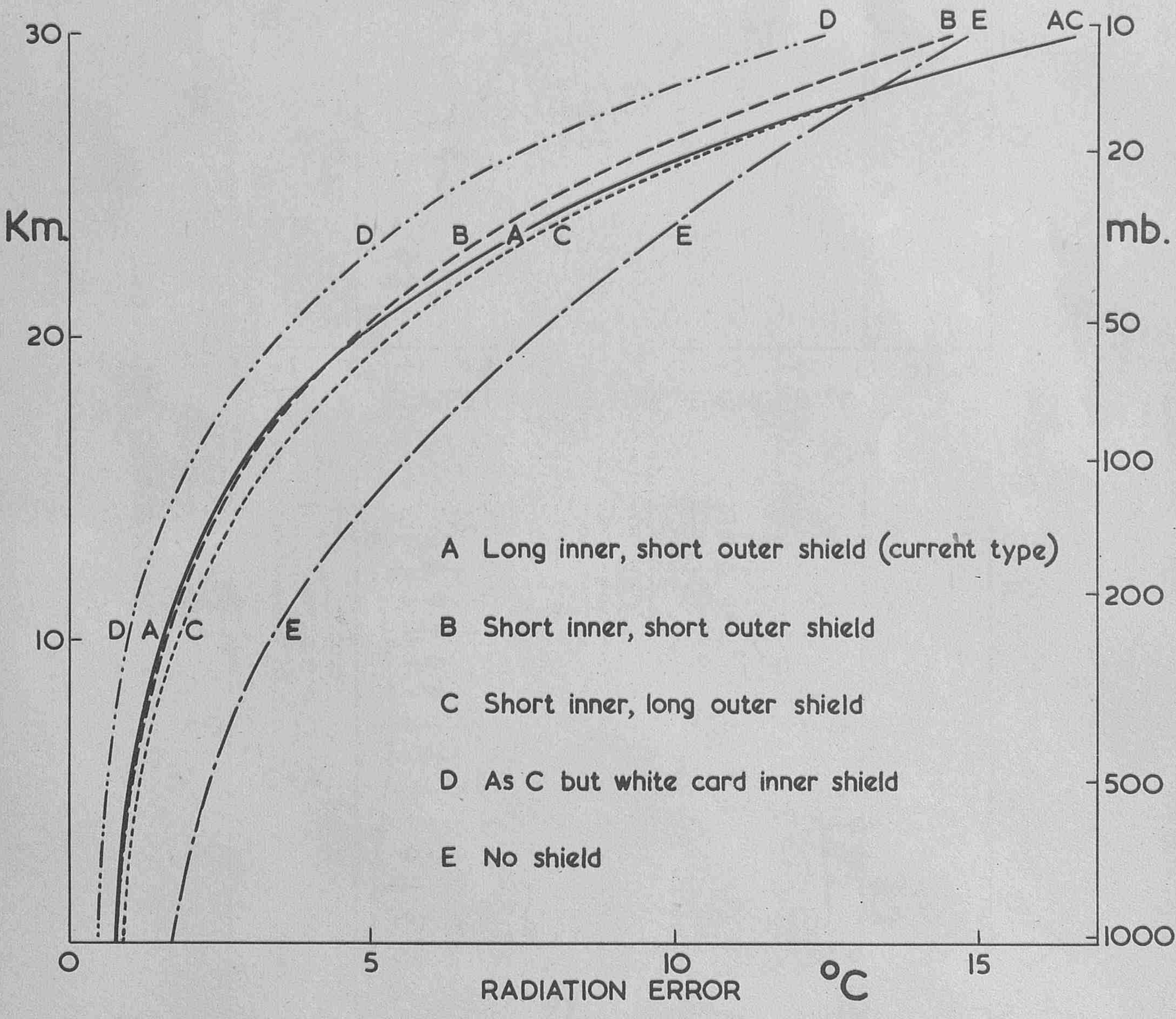


Fig. 4 Calculated radiation errors with standard and modified shields (solar altitude 40°)

M.R.P. 836

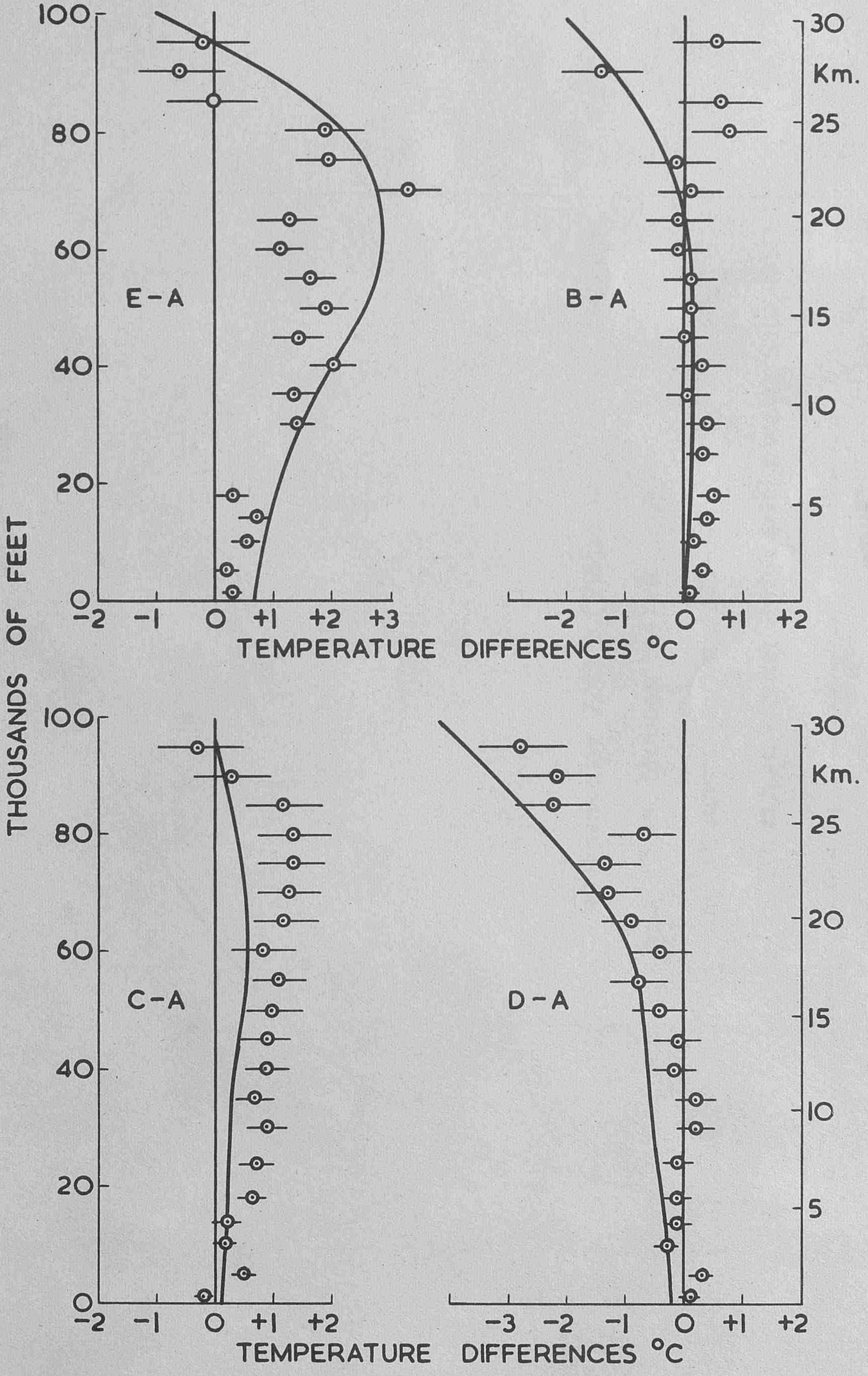


Fig. 5. Observed and calculated temperature differences with standard and modified shields (solar altitude 40°)

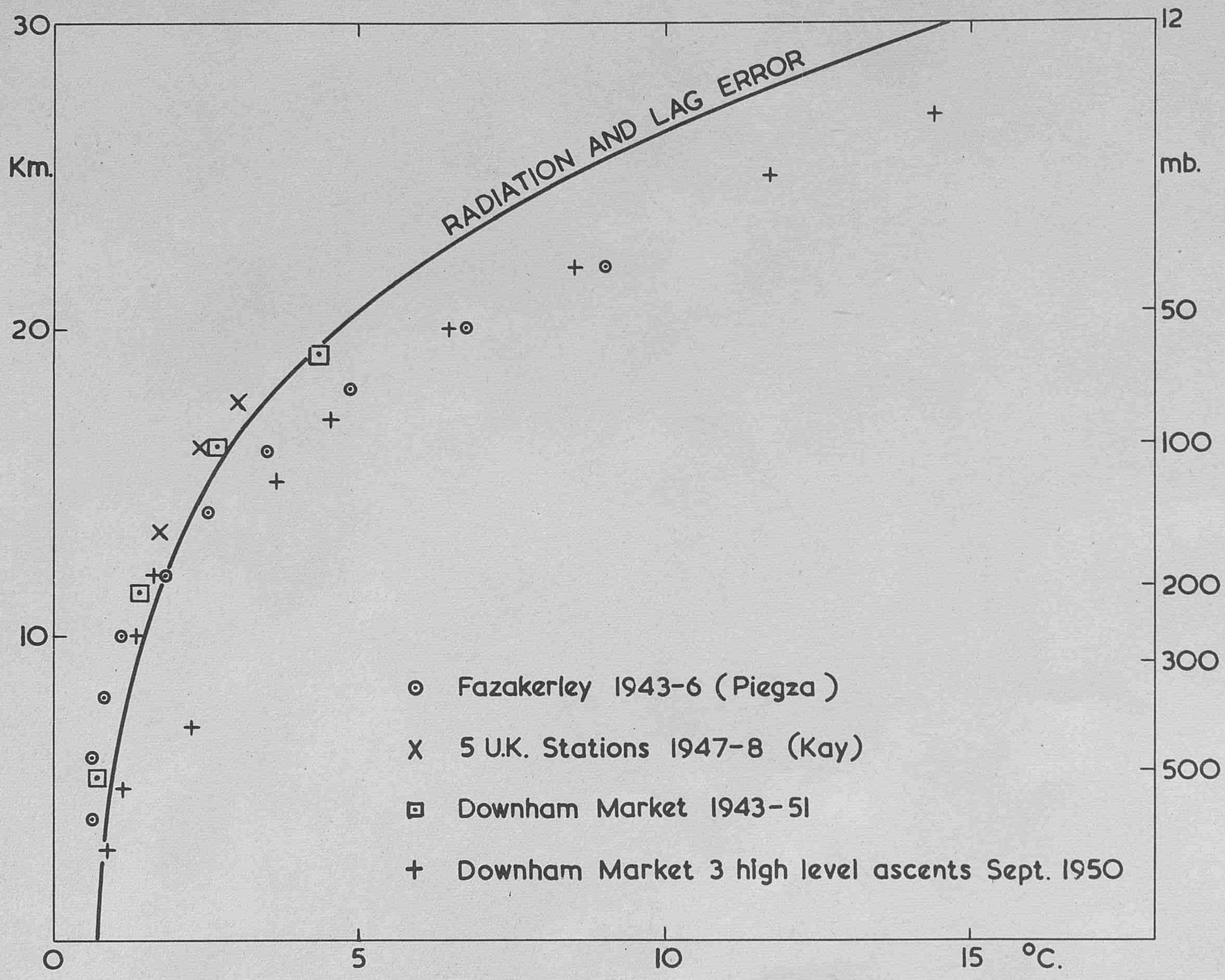


Fig. 6 Day and night temperature differences compared with instrument errors (solar altitude about 40°)

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S.C.II/160

20th October, 1953.

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[illegible] have occurred there. It was [illegible] boundaries for all purposes [illegible] at some time during their existence, and in order to obtain accurate statistics of their duration for example it was necessary to follow them until they finally disappeared from the charts.

Thus in this report the following convention was adopted. Those statistics relating to the duration of pools (sections 1 and 2) apply only to those which passed at least a half of their total existence within the area in question, part of their lives (the lesser half) being spent outside in a minority of cases. Those statistics of more "local" significance, relating to cold pools en masse (sections 3 and 4), apply to all the pools within the area and to none other.

Another distinction observed here (in sections 1, 2 and 4) is between those pools which at some time during their existence reached intensity two or more* and those which were at the minimum intensity throughout.

1. The more intense pools. In M.R.P. 764 only those pools contained within two or more closed thickness lines and appearing on at least two successive 0300 G.M.T. circumpolar thickness charts, were considered. These 75 pools were often extended before and/or after by periods during which they were weaker, i.e. of intensity one, and were on occasions followed outside the specified area. The above requirement that the pools must spend not less than half their total life within the area, reduces their number considered herein to 54. To these may be added 9 cases associated with them (or with larger pools outside the area) as subsidiary features, i.e., contained within a common outer closed thickness line but lasting only a day, and 111 which appeared on only one 0300 G.M.T. chart at intensity two or more. The latter however usually lasted beyond a day as weaker pools.

At the time of issue this report has not been published. Any author desiring to quote this paper should refer to it as "A paper of the Meteorological Research Committee (London),
M.R.P. No. 837 by Mr. E. J. Sumner
A copy is available in the Library of the Meteorological Office."

M.R.P. 837

S.C.II/160

20th October, 1953.

AD 63909

## AIR MINISTRY
## METEOROLOGICAL RESEARCH COMMITTEE

### The annual and geographical distribution of cold pools
by
E.J. Sumner (Meteorological Office)

A previous study has been made (M.R.P. 764) of the more intense cold pools of the Atlantic-European sector of the Northern Hemisphere, for the 5-year period September, 1946 to August, 1951, inclusive. The present report deals with all the cold pools (areas within one or more closed lines in the 1000-500 mb. thickness topography and occurring on an 0300 G.M.T. chart) within the same region and for an almost identical period, viz., October, 1946 to September, 1951, inclusive. The precise area considered was bordered to the west by the 60°W meridian and to the east by the 30°E meridian. Pools north of 80°N and south of 30°N were omitted because the observational coverage was inadequate to locate all the pools which might have occurred there. It was however impossible to adhere rigidly to these boundaries for all purposes since some of the pools traversed them at some time during their existence, and in order to obtain accurate statistics of their duration for example it was necessary to follow them until they finally disappeared from the charts.

Thus in this report the following convention was adopted. Those statistics relating to the duration of pools (sections 1 and 2) apply only to those which passed at least a half of their total existence within the area in question, part of their lives (the lesser half) being spent outside in a minority of cases. Those statistics of more "local" significance, relating to cold pools en masse (sections 3 and 4), apply to all the pools within the area and to none other.

Another distinction observed here (in sections 1, 2 and 4) is between those pools which at some time during their existence reached intensity two or more* and those which were at the minimum intensity throughout.

1. The more intense pools. In M.R.P. 764 only those pools contained within two or more closed thickness lines and appearing on at least two successive 0300 G.M.T. circumpolar thickness charts, were considered. These 75 pools were often extended before and/or after by periods during which they were weaker, i.e. of intensity one, and were on occasions followed outside the specified area. The above requirement that the pools must spend not less than half their total life within the area, reduces their number considered herein to 54. To these may be added 9 cases associated with them (or with larger pools outside the area) as subsidiary features, i.e., contained within a common outer closed thickness line but lasting only a day, and 111 which appeared on only one 0300 G.M.T. chart at intensity two or more. The latter however usually lasted beyond a day as weaker pools.

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* The intensity of a cold pool is defined in accordance with the convention: one closed thickness line, intensity one; two closed lines, intensity two; ........ and so on. The isopleths are drawn at intervals of 200 ft. (even hundreds) on the circumpolar charts.

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The frequency distribution of the duration of these 174 more intense pools (viz., those which at some time or other were of intensity two or more) is shown in Fig. 1. For each spell duration, the number of pools of the minimum intensity is indicated together with the 'internal' distribution of periods (usually continuous) at the higher intensities (two or more). The average duration of these pools is 4.0 days with a mode of 3-4 days. One-day pools are very infrequent, so that a forecaster can be fairly certain that an intense pool will still be present 24 hours after its first appearance.

2. The weaker pools. The remaining pools are necessarily of intensity one throughout; their frequency distribution is shown in Fig. 2. These weaker pools were usually quite ephemeral, most frequently lasting only a day, the average duration being 1.6 days.

3. Central thickness anomalies. A measurement of the 1000-500 mb. thickness at the centre of each pool was made, by interpolation to the nearest 50 ft. The frequency distribution of central thickness anomaly (departure from a 5-year mean for the appropriate month) is shown in Fig. 3, the pools now being segregated simply in accordance with their intensity. The more intense pools (intensity two or more) always had a central thickness below the average for the time and place, their average anomaly being -570 ft. The intensity-one pools, too, almost invariably had a central thickness below average but with a higher general level of thickness, the average anomaly being -345 ft.

4. Annual and geographical distribution. The distribution by months throughout the year of the more intense and the weaker pools is shown in Figs. 4 and 5, respectively, each occurrence on an 0300 chart now being counted. These are largely the pools considered in sections 1 and 2, but the totals are not the same because of the convention described in the introduction, which now excludes the pools which passed outside the area.

The occurrences of the more intense pools were most frequent in May and June and least frequent in August and again in January, as previously demonstrated in M.R.P. 764 for somewhat similar data. The weaker pools, on the other hand, have a very dissimilar distribution, with a pronounced maximum in August.

There were often several pools simultaneously present on an 0300 G.M.T. chart: on about 55% of occasions there was only a single pool, with two pools per chart on about 30% and three pools on 12%; the remaining 3% had four and (rarely) five pools per chart. There were less pools per chart in winter than summer, the respective figures for charts with a single pool being about 70% and 40%.

The monthly mean thickness anomalies for each of the above classes are shown in Table I.

Table I. Monthly mean thickness anomalies: tens of feet

| Pool within type of spell | Jan. | Feb. | Mar. | Apr. | May | June | July | Aug. | Sept. | Oct. | Nov. | Dec. | Year |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| More intense | 62.5 | 47.6 | 54.1 | 50.8 | 40.6 | 40.1 | 39.3 | 37.2 | 41.3 | 42.1 | 35.1 | 60.0 | 42.6 |
| Weaker | 31.3 | 36.1 | 33.5 | 34.9 | 29.2 | 34.1 | 30.9 | 27.2 | 29.6 | 30.4 | 34.0 | 31.3 | 31.3 |

For the pools of the weaker spells there is little change in the mean anomaly from month to month, whereas for the more intense pools there is a distinct seasonal variation.

The positions of all the pools (occurrences at 0300 hours) within the area have been plotted on charts, a month at a time (Figs. 6-17). Pools of intensity two or more are distinguished from lesser pools. The main features of their geographical distribution are as follows:-

(1) From October to March the pools are mainly concentrated over Europe and in the Mediterranean. Within this period, there are also several regions with either no pools at all in some months or a very sparse coverage in others. The most noticeable of these are: the area from Greenland eastwards to the Norwegian Sea in latitudes 60-70°N, and the region of the Atlantic in about 40-50°N extending eastwards from the American seaboard to as far as 10°W in January but less in other months.

(2) From April to September cold pools are more uniformly distributed over the area considered, although still thinnest in the regions mentioned in the previous paragraph. Also they become very infrequent in the Mediterranean in July and August.

(3) There is a marked scarcity throughout the year in the extreme south-western corner of the area. This may however be partly due to the small number of upper-air observations there.

(4) Areas subject to cold pools usually have both classes of intensity. One exception is, however, the area roughly west and north-west of the British Isles (extending to 30-40°W some months) where the more intense pools are practically non-existent, or if they do occur quickly weaken. A synopsis of all the data, on a seasonal basis, is contained in Fig. 18.

5. Comments. As has been remarked by C.K.M. Douglas (1947), cold pools are most frequent when and where the normal latitudinal temperature gradient is weak. The annual distribution of the weaker pools (Fig. 5) is certainly in keeping with this principle. Moreover, an inspection of the monthly mean thickness charts for the corresponding 5-year period, revealed that for all the pools the greatest areal concentrations each month also occurred where the mean thermal gradients were weak. The converse is not, however, always true; for example, the winter minimum of pools over Iceland is in an area for which the thermal gradient is as weak as over Europe, where the maximum concentration occurred. The (inverse) relationship between the mean thermal gradient and the number of cold pools is not of course a direct causative one: the reverse effect, viz., the more pools that occur the weaker the mean thermal gradient becomes, must operate to a large extent.

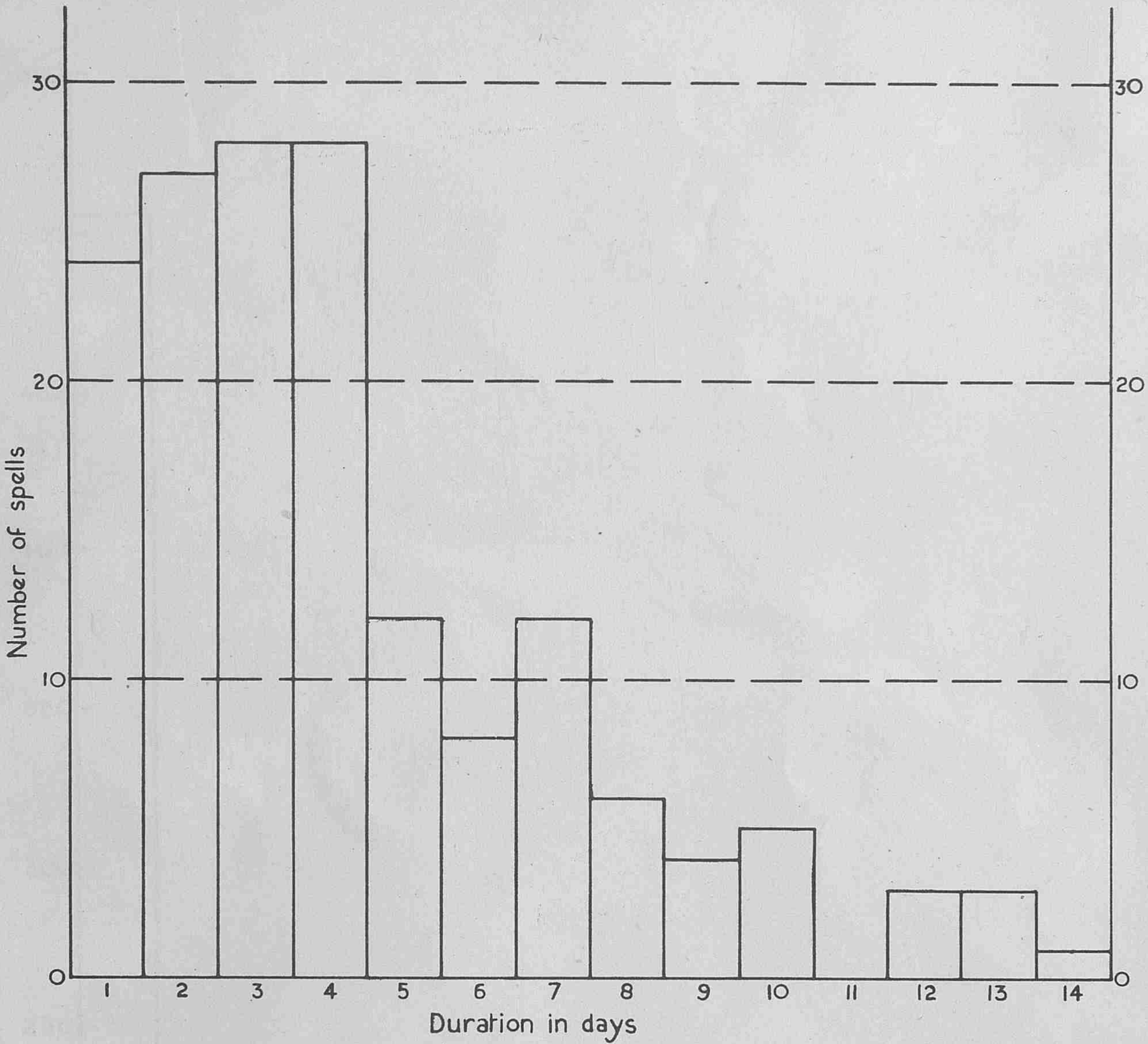
Although the maximum monthly concentrations of the pools occurred in association with weak troughs in the monthly mean thickness topography, there were (lesser) concentrations associated with weak ridges, e.g., that over and to the west of the British Isles in February.

The seasonal distribution of the more intense cold pools is rather difficult to explain. One might possibly expect a late spring and a late autumn optimum on the grounds that the mean thermal gradient is too strong in winter and the heating too intense in summer, to allow the formation of large pockets of relatively cold air in the Atlantic and European sectors. The mean figures in Table I suggest how much colder the air has to be in winter in order to establish an intense cold pool against the strong prevailing thermal gradients. In summer, on the other hand, a much smaller anomaly would suffice to give an intensity-two pool but this is evidently seldom achieved in face of the (non-adiabatic) warming.

The enhanced late spring and early summer peak, and the relative "failure" of the late autumn maximum, must be connected with the seasonal warming and cooling, respectively, and with the synoptic systems which are associated with these processes. The association of the more intense cold pools with blocking, which has a similar seasonal distribution in the same sector (see Bretzowsky et alia, 1951, for example), is probably significant in this connection.

## References

| | | |
|---|---|---|
| Bretzowsky, H., H. Flohn and P. Hess | - | Some remarks on the climatology of blocking action. Tellus, Vol. 3, No. 3, 1951. |
| Douglas, C.K.M. | - | Cold pools. Met. Mag., Vol. 76, No. 903, 1947. |



The figures given below represent the number of occasions on which pools of a given spell duration (reading across) were of intensity two or more for (reading down) one day, two days - - - etc., respectively (usually but not necessarily consecutively)

| No. of days at intensity two or more | Length of spell in days<br>1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 24 | 25 | 24 | 13 | 5 | 1 | 4 | 2 | 1 | 1 | — | — | — | — |
| 2 |  | 2 | 4 | 11 | 6 | 4 | 5 | 2 | — | 1 | — | 1 | 1 | — |
| 3 |  |  | — | 4 | 1 | — | 1 | — | 1 | 1 | — | — | 1 | — |
| 4 |  |  |  | — | — | 1 | 1 | 1 | 1 | 2 | — | 1 | 1 | — |
| 5 |  |  |  |  | — | 1 | 1 | — | 1 | — | — | 1 | — | — |
| 6 |  |  |  |  |  | 1 | — | — | — | — | — | — | — | — |
| 7 |  |  |  |  |  |  | — | 1 | — | — | — | — | — | — |
| 8 |  |  |  |  |  |  |  | — | — | — | — | — | — | — |
| 9 |  |  |  |  |  |  |  |  | — | — | — | — | — | 1 |

Occurrences of pools on 0300 z charts

|  |  |  |  |  |  |  |  |  |  |  |  |  |  |  |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Total no. of pools of higher intensity (≥2) | 24 | 29 | 32 | 47 | 20 | 24 | 26 | 17 | 13 | 14 | — | 11 | 9 | 9 |
| Total no. of intensity one | 0 | 25 | 52 | 65 | 40 | 24 | 58 | 31 | 23 | 36 | — | 25 | 30 | 5 |

Fig. 1 The more intense cold pools: Frequency distribution of spell-duration October 1946 – September 1951

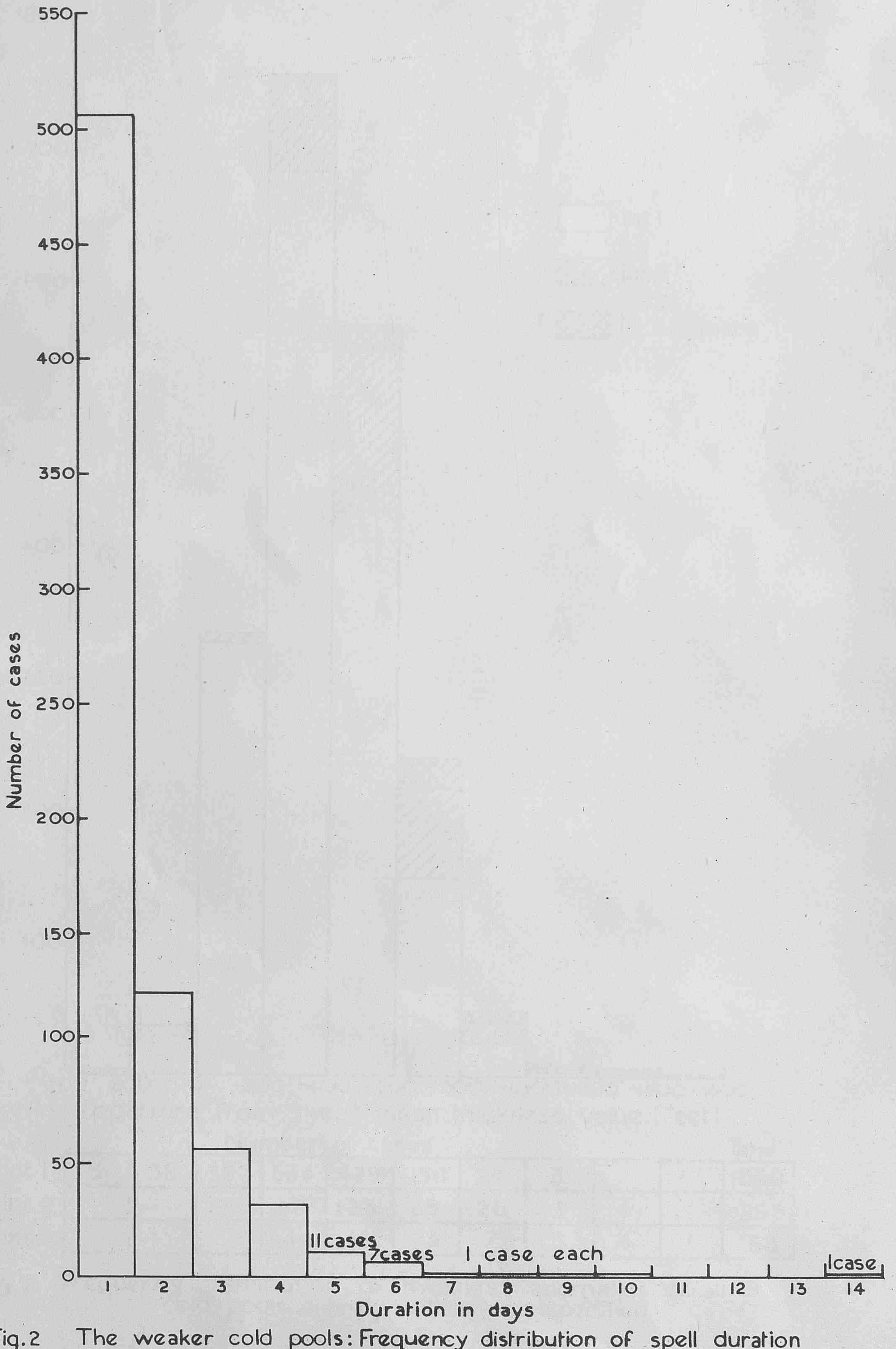


Fig. 2 The weaker cold pools: Frequency distribution of spell duration October 1946 – September 1951

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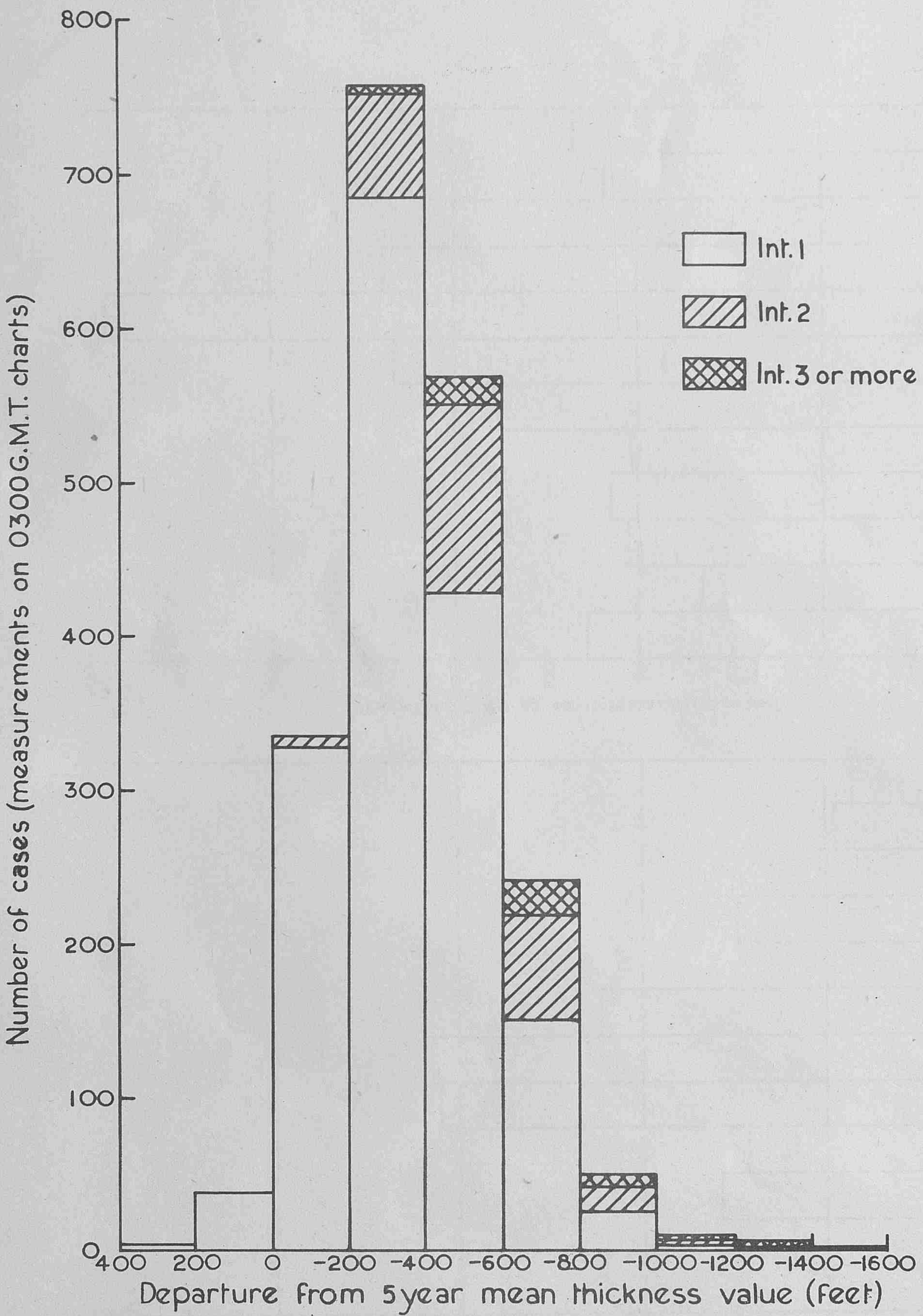


Number of cases

| | | | | | | | | | | | Total |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Int. 1 | 3 | 38 | 327 | 684 | 429 | 150 | 26 | 3 | | | 1660 |
| Int. 2 | | | 7 | 68 | 123 | 68 | 26 | 3 | 1 | | 296 |
| Int. 3 | | | | 4 | 16 | 23 | 7 | 3 | 4 | 1 | 58 |

Fig. 3 Frequency distribution of thickness anomalies of all cold pools entirely within area specified

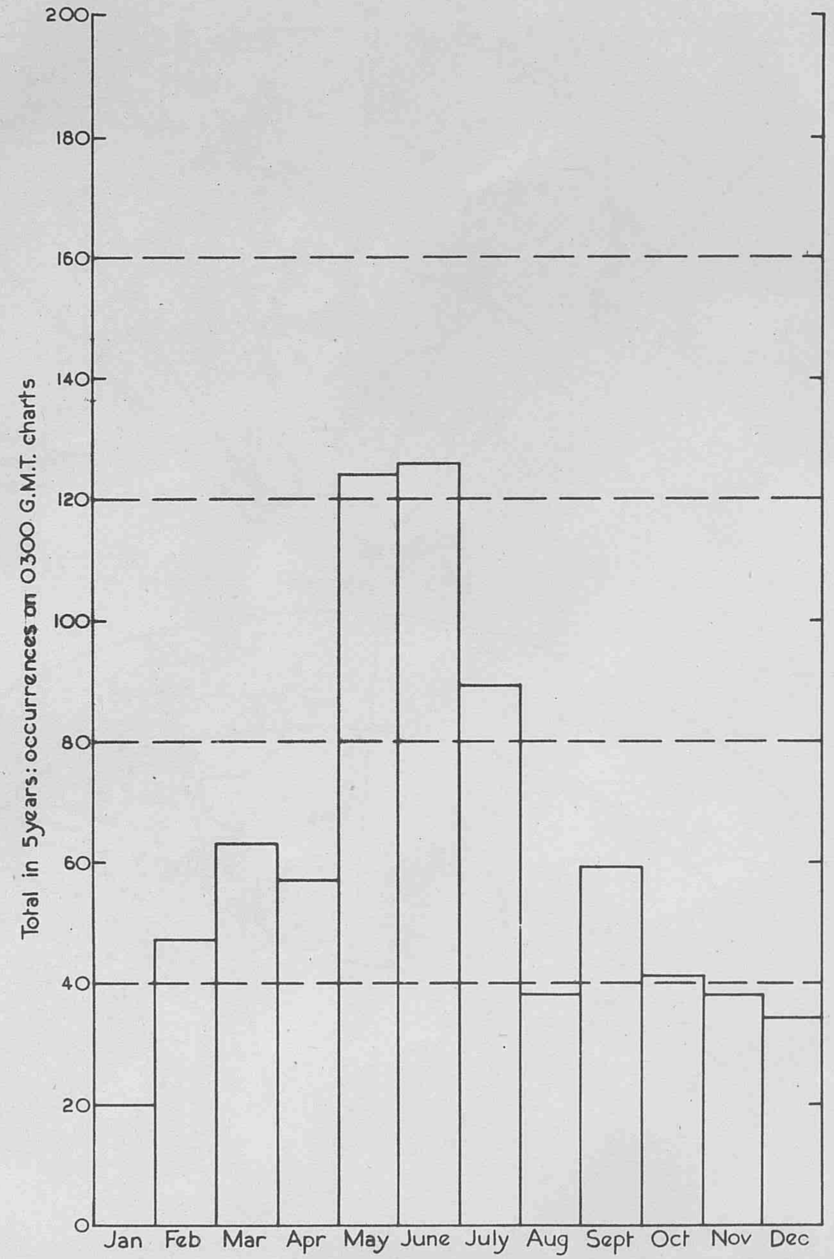


Fig.4 Histogram of monthly distribution of the more intense pools (Each occurrence on an 0300Z chart counted once)

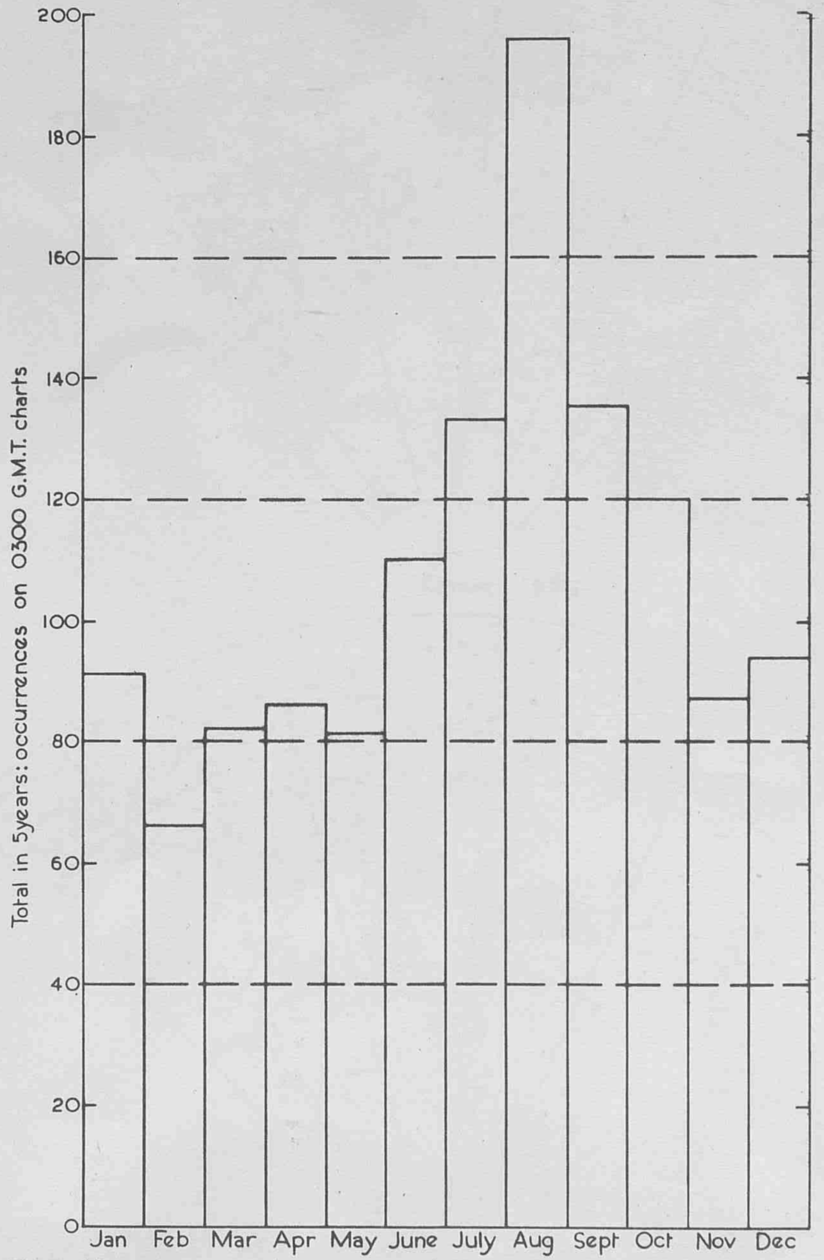


Fig.5 Histogram of monthly distribution of the weaker pools

# Cold Pools: October 1946 – September 1951

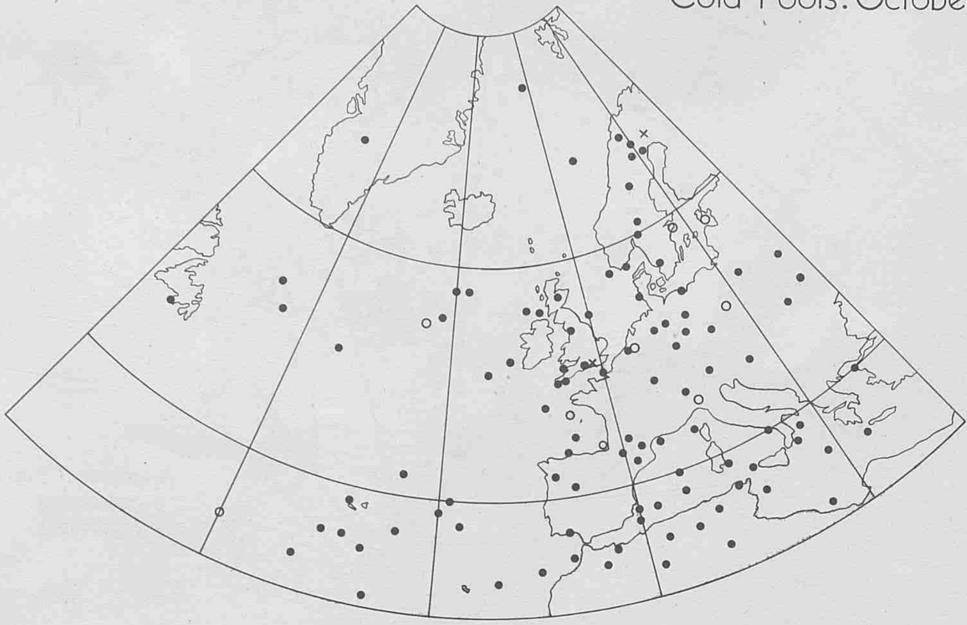


Fig. 6 January

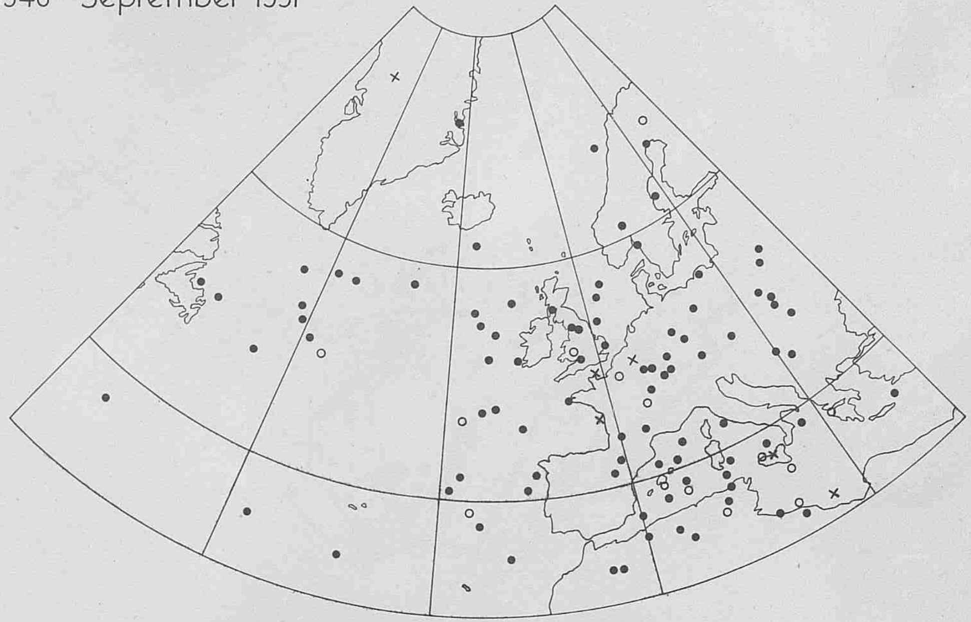


Fig. 7 February

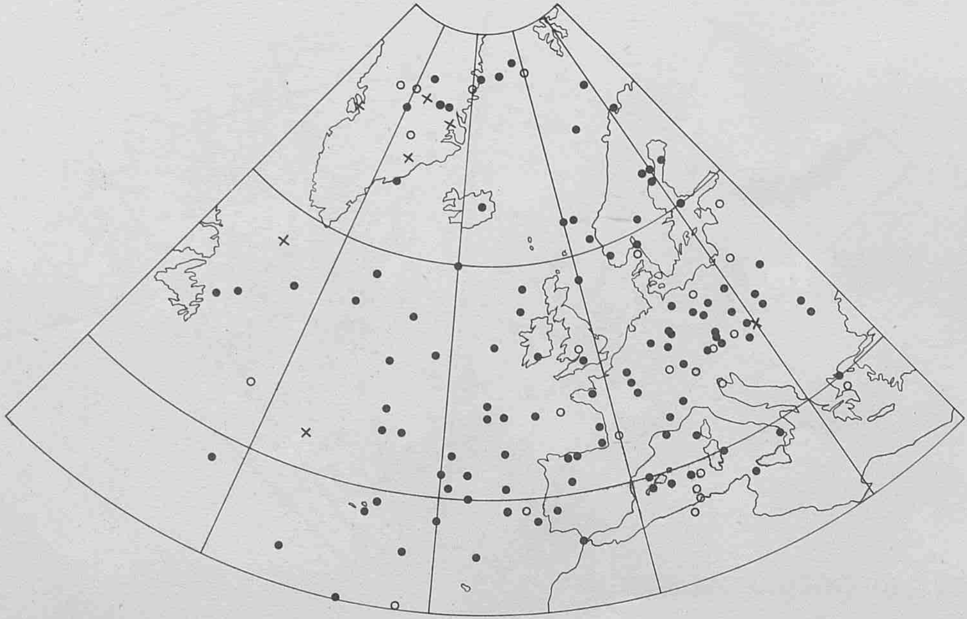


Fig. 8 March

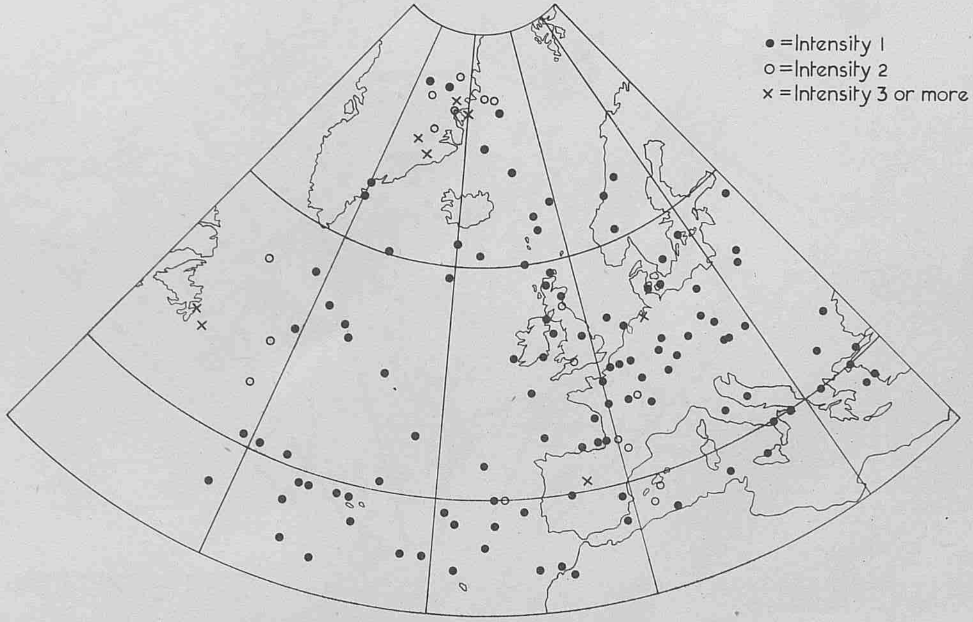


Fig. 9 April

● = Intensity 1
○ = Intensity 2
× = Intensity 3 or more

# Cold Pools: October 1946–September 1951

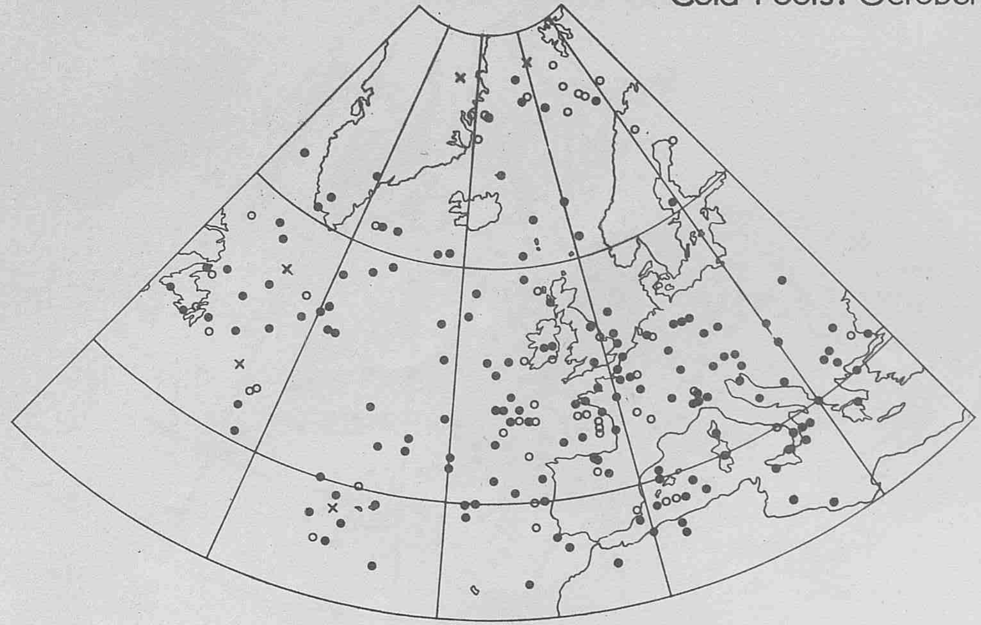


Fig. 10 May

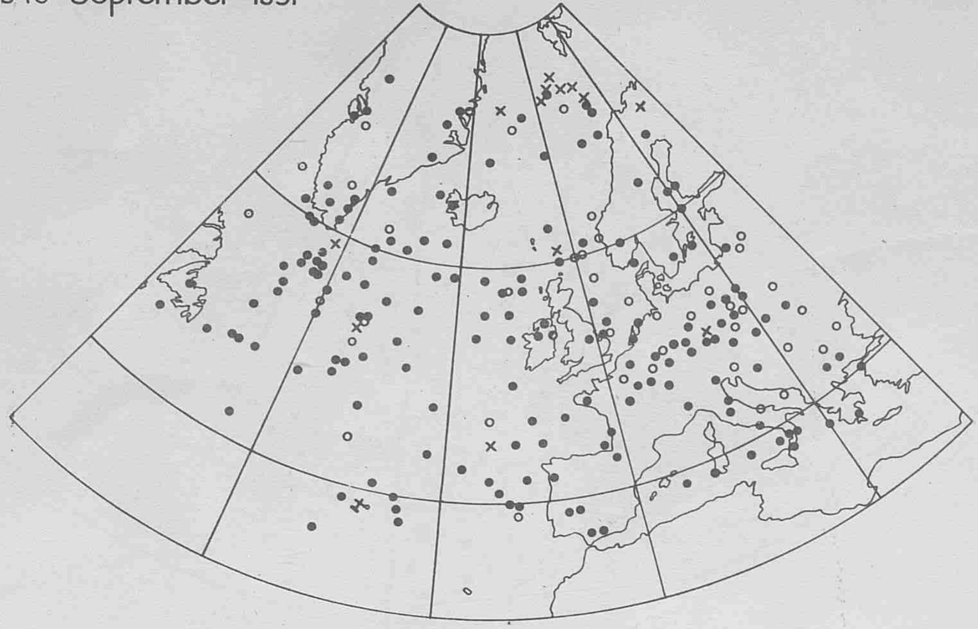


Fig. 11 June

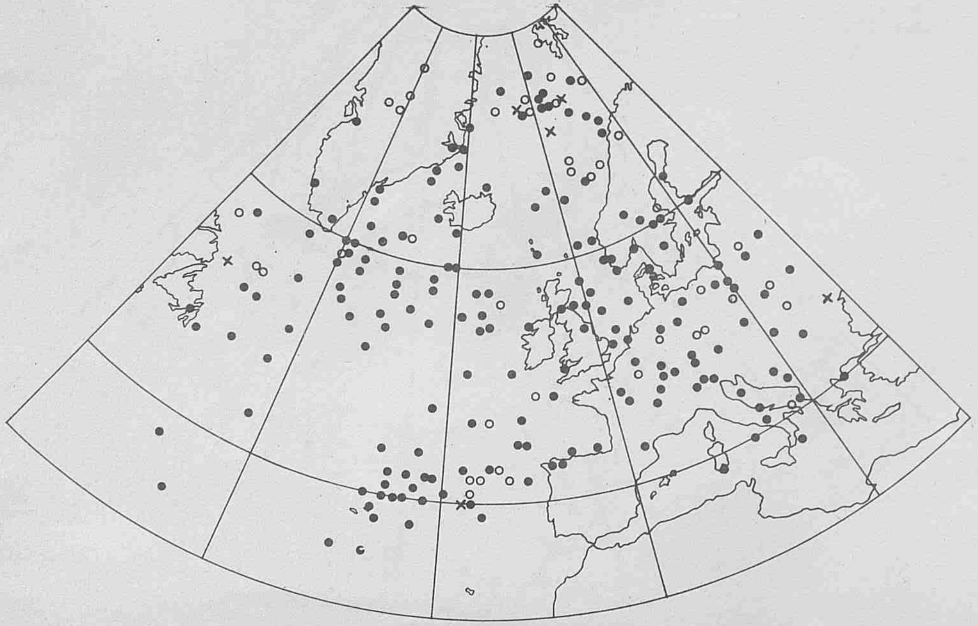


Fig. 12 July

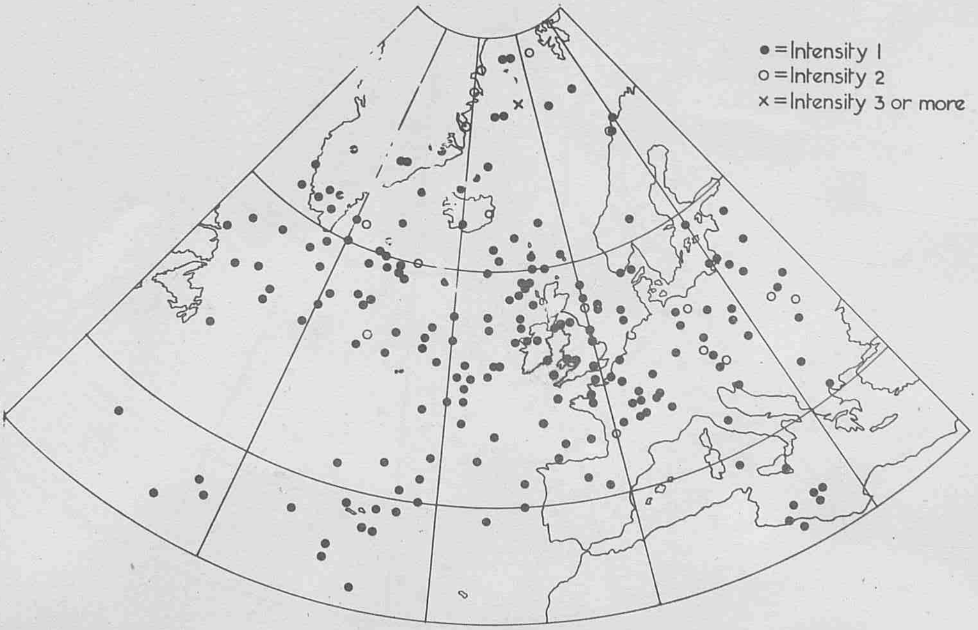


Fig. 13 August

● = Intensity 1
○ = Intensity 2
× = Intensity 3 or more

# Cold Pools: October 1946–September 1951

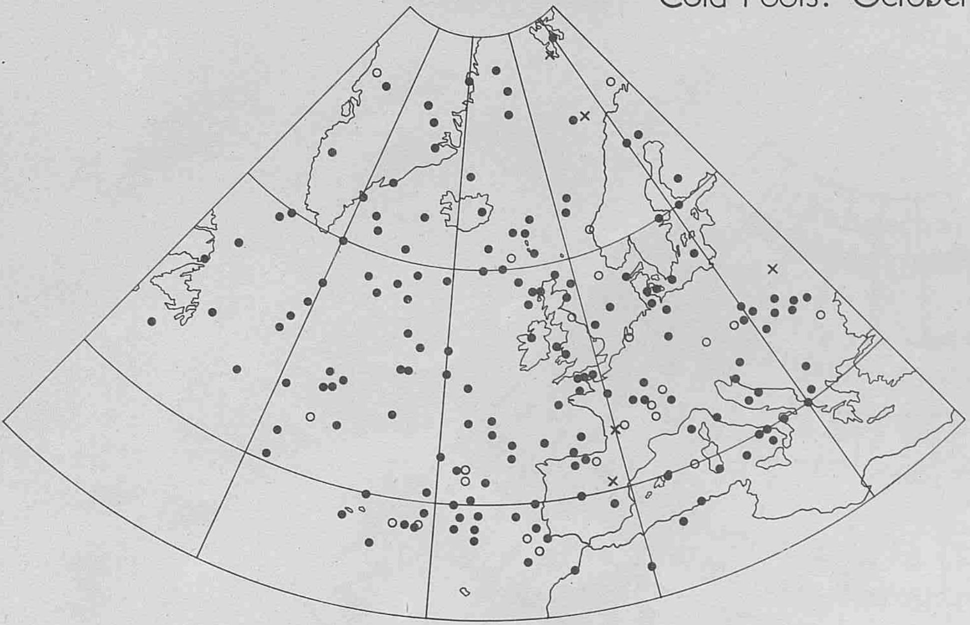


Fig. 14 September

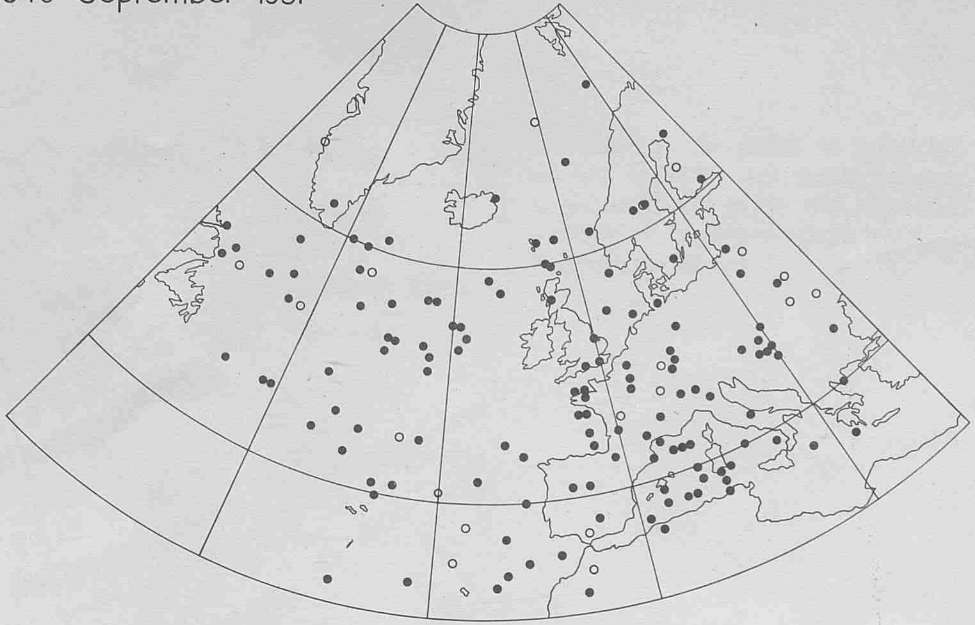


Fig. 15 October

● = Intensity 1
○ = Intensity 2
× = Intensity 3 or more

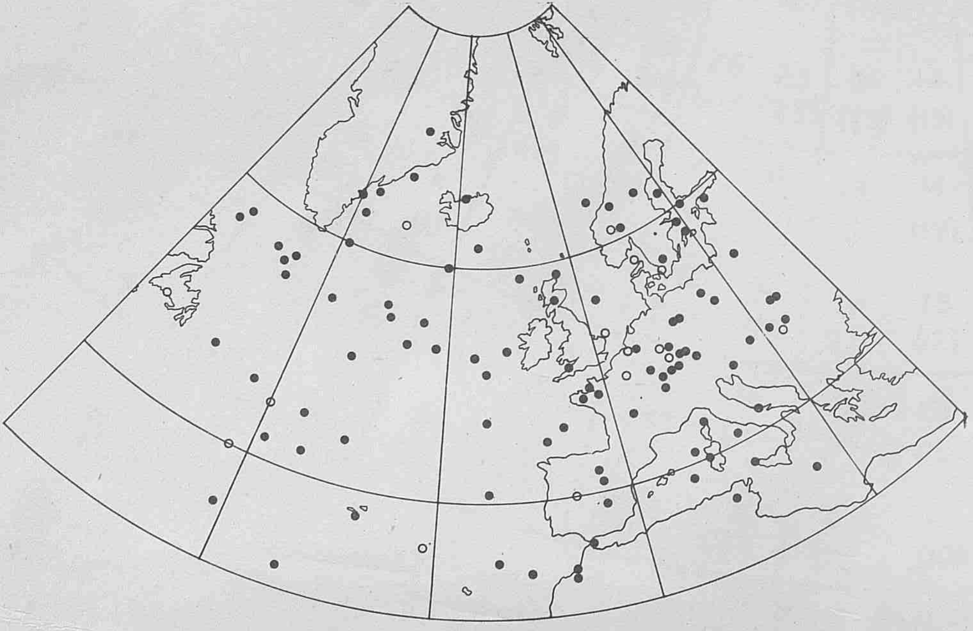


Fig. 16 November

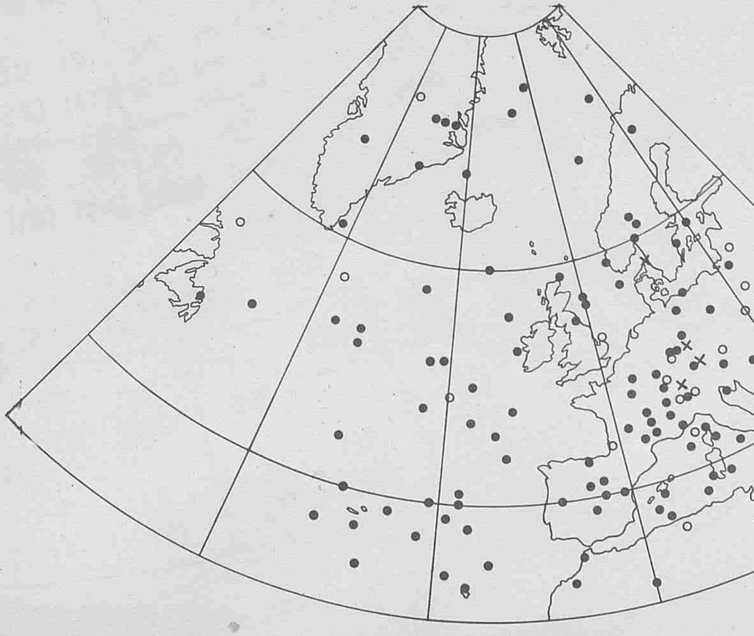
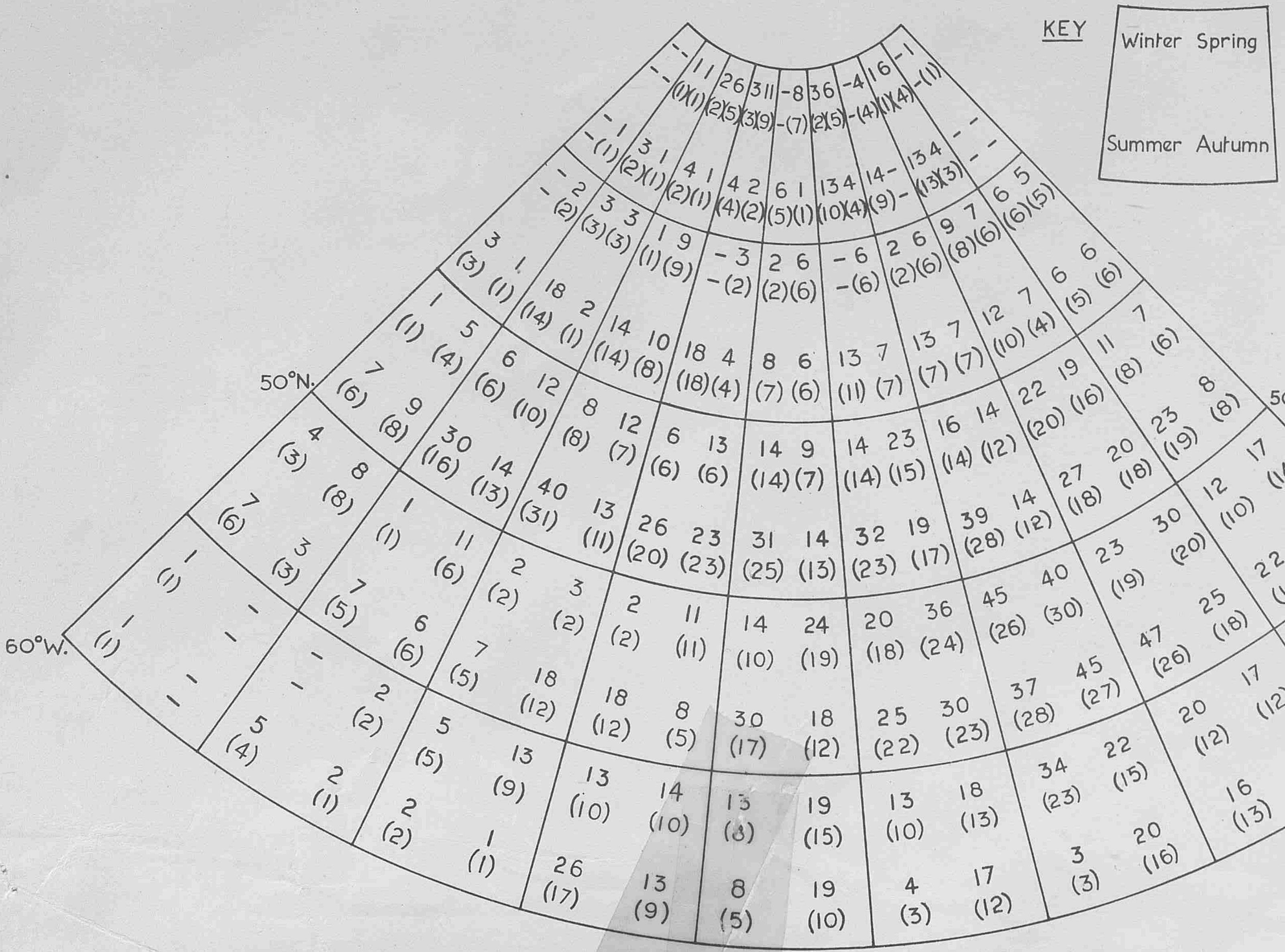


Fig. 17 December


KEY
Winter Spring
Summer Autumn
60°W.
50°N.
50°N.
30°E.
– – – –
1 1 (1)(1)
2 6 (2)(5)
3 11 (3)(9)
– 8 – (7)
3 6 (2)(5)
– 4 – (4)
1 6 (1)(4)
– 1 – (1)
– 1 – (1)
3 1 (2)(1)
4 1 (2)(1)
4 2 (4)(2)
6 1 (5)(1)
13 4 (10)(4)
14 – (9) –
13 4 (13)(3)
– – – –
– 2 – (2)
3 3 (3)(3)
1 9 (1)(9)
– 3 – (2)
2 6 (2)(6)
– 6 – (6)
2 6 (2)(6)
9 7 (8)(6)
6 5 (6)(5)
3 1 (3)(1)
18 2 (14)(1)
14 10 (14)(8)
18 4 (18)(4)
8 6 (7)(6)
13 7 (11)(7)
13 7 (7)(7)
12 7 (10)(4)
6 6 (5)(6)
1 5 (1)(4)
6 12 (6)(10)
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16 14 (14)(12)
22 19 (20)(16)
11 7 (8)(6)
7 9 (6)(8)
30 14 (16)(13)
40 13 (31)(11)
26 23 (20)(23)
31 14 (25)(13)
32 19 (23)(17)
39 14 (28)(12)
27 20 (18)(18)
23 8 (19)(8)
4 8 (3)(8)
7 3 (6)(3)
1 11 (1)(6)
2 3 (2)(2)
2 11 (2)(11)
14 24 (10)(19)
20 36 (18)(24)
45 40 (26)(30)
23 30 (19)(20)
12 17 (10)(14)
– – (1) –
– – 5 (4)
7 6 (5)(6)
– 2 (2)
2 (1)
7 18 (5)(12)
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2 1 (2)(1)
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8 19 (5)(10)
25 30 (22)(23)
13 18 (10)(13)
4 17 (3)(12)
37 45 (28)(27)
34 22 (23)(15)
3 20 (3)(16)
47 25 (26)(18)
20 17 (12)(12)
16 13 (13)(10)
22 26 (14)(21)
7 9 (6)(6)
5 4 (4)(3)

The number of pools in each 10° square is given for each season in accordance with the key: the number of different spells involved is given in brackets

Fig. 18 Distribution of cold pools
October 1946–September 1951

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Committee (London), M.R.P. 853 by
Mr. .................................. A copy is available in
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M.R.P. 853
27th April, 1954

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## AIR MINISTRY

## METEOROLOGICAL RESEARCH COMMITTEE

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M.R.P. 860
S.C.II/167
27th March, 1954

# AIR MINISTRY

# METEOROLOGICAL RESEARCH COMMITTEE

## Variation of temperature in the lower stratosphere above Larkhill and Lerwick

(Addendum to M.R.P. 819)

by

J.K. Bannon (Meteorological Office)

### Introduction and Summary

The numbers of observations for 60 mb. used in the analysis described in the previous paper, M.R.P. 819(1), were not large. Moreover, data for all hours of observation, day and night, were used. Because of the comparatively scanty data and the possibility that radiation errors might result in correlation coefficients being erroneously small, the stratospheric correlation coefficients were computed again for Larkhill and Lerwick using all available data for the level of 60 mb. for specified hours of observation, day and night separately. The important result obtained previously, namely that the relation between temperature and the height of the 300 mb. surface and tropopause pressure respectively, falls off rapidly between 150 and 60 mb. was confirmed, but the seasonal variations in these relations appear to be less marked than was indicated in the earlier paper.

### 1. Data

As in M.R.P. 819 the analysis was restricted to data for the four mid-season months, January, April, July and October. All observations reaching the level of 60 mb. at 00 and 12h. G.M.T. before 1948 and at 03 and 15h. G.M.T. in 1948 and later years at Larkhill and Lerwick were used and values of the height of the 300 mb. surface ($H_{300}$), the pressure of the tropopause ($P_T$), and temperatures at levels 150, 100 and 60 mb. ($T_{150}$, $T_{100}$, $T_{60}$ respectively) were extracted, the day and night observations being treated separately. For January and April at Lerwick and April at Larkhill so few observations were available for the selected hours that it was necessary to use data for other hours of observation, the data being grouped according to whether the observations were in daylight or darkness. These data provided comparable sets of observations at each of the three levels, 150, 100 and 60 mb. in contrast with the data used in M.R.P. 819 which embraced many more observations at 150 mb. than at 60 mb. whence there was always the possibility that observations at the higher level, through some bias or another, might not be in the same class as the observations from lower levels.

Correlation coefficients between $H_{300}$ and $T_{150}$, $T_{100}$ and $T_{60}$ respectively and between $P_T$ and the same temperatures were computed. The results are given in Tables I and II for Larkhill and Lerwick respectively. Results for day and

/night

night are given separately although in several months the figures for one or other of day or night have no great significance. They serve, however, as a useful reminder of the limitations of correlation analysis and help in assessing the accuracy of the results.

## 2. Discussion

### (a) Larkhill. Table I.

The day and night sets appear to be consistent and there is no significant difference in the correlation coefficients from those given in Table I of M.R.P. 819 except that values at 60 mb. are generally smaller than in M.R.P. 819.

As a means of judging whether the selected observations used in the preparation of Table I are representative of general conditions in the stratosphere, Table III was prepared comparing some of the statistics in Table I with similar statistics for all observations in the period 1946-1950. In some months the standard deviations differ appreciably but considering the numbers of observations it can be said that the selected sets are representative of general conditions. The majority of the day/night observations used in compiling Table I were separated by more than 24 hours in time. Persistence effects will thus be small and the effective number of observations must be little short of the actual number used. (See discussion in M.R.P. 819).

### (b) Lerwick. Table II.

Here there are significant differences between the night and day except in October. In January the day correlation coefficients show weaker relationships than the night at each of the stratospheric levels. Table IV, prepared as Table II to compare the sets of selected observations with the more general observations, shows that for the day observations the pressure of the tropopause was much less variable and the temperature at 150 mb. slightly less variable than in general. The night set with nearly double the number of observations is much more in agreement with the general unselected data. It is deduced that the day coefficients are not representative and that the night coefficients are more likely to represent the general relations. The differences between the day and night correlation coefficients in April and July are also almost certainly not significant as the number of night observations is small in each case; in July there is small variability in $P_T$ and $T_{150}$ for the night observations.

To get a more representative set of correlation coefficients for January and to reduce sampling errors the day and night observations were combined. The resulting statistics are given in Table V. Data are sufficient to make only a rough estimate of diurnal variation of temperature at 60 mb. as observed by radio-sonde at Lerwick in January. The statistics in Table II would suggest a figure of 2-3°F. A comparable figure for Larkhill would be 4-5°F. from the data for 1946-1950(2) but rather less (3-4°F.) from the figures in Table I assuming that diurnal variation at 150 mb. is of the order 1°F. As discussed in M.R.P. 819 such a small variation between day and night observations would have a very small effect on correlation coefficients computed from these observations combined when the standard deviation of temperature is three or four times as large as the semi-diurnal range.

The correlations for January in Table V and for daylight for April, July and October in Table II do not show great differences from those in Table II of M.R.P. 819. The other statistics in Tables II and V compared with the statistics for unselected ascents in Table IV show that these selected data for high altitude are fairly representative of general conditions.

The effective number of observations must, as in Table I, be only a few less than the number of observations.

/3.

## 3. Variation of the correlation coefficients with height

The correlation coefficients in Tables I, II and V show a marked decrease with decreasing pressure. The direct relationship between two variables varies as the square of their correlation coefficient. Accordingly Table VI was prepared giving the ratio of the squares of the correlation coefficients for $T_{60}$ and $T_{150}$ (compare Table VII, M.R.P. 819). It is a reasonable inference from these figures that the influence of the disturbances of the troposphere in middle latitudes on temperature in the stratosphere is reduced to at least a half between 150 and 60 mb. and for much of the year to one-fifth or one-third.

## 4. Variation of the correlation coefficients with season

At Larkhill (Table I) the correlation coefficients are lowest in January and highest in July and October. In M.R.P. 819 the coefficients for January are almost as great as those for October. Except at 60 mb. the differences between the seasons are not large. Table VI shows that the decrease with height in the relationship between stratospheric temperatures and $H_{300}$ and $P_T$ respectively is least in October and greatest in January and April.

At Lerwick (Tables II and V) the correlation coefficients are least in January, not very much greater in April and largest in October. January and October have the greatest and least reductions respectively in the coefficients between the 150 and 60 mb. levels (Table VI). The greatest correlation coefficients (October) have similar values at Larkhill and Lerwick but the coefficients at Lerwick are lower than at Larkhill in January and April especially at 60 mb. where the differences between the coefficients for $H_{300}$ are certainly significant. The same types of differences were noted in M.R.P. 819.

It was suggested in M.R.P. 819 that the variability in temperature distribution over Lerwick in January, where two different stratospheric regimes can occur, is the cause of the lower correlations in that month. This argument has some force. It is noted, however, that as for $H_{400}$ at Arctic Bay (M.R.P. 819, Table III), it is the $H_{300}$ correlations which are markedly reduced, the correlations involving $P_T$ being not much less than in summer. Much of the relation between pressure at the tropopause and stratospheric temperature must arise from the close relation between vertical motion at the tropopause and vertical motion within the lower stratosphere; a low tropopause can be partly caused by subsidence and this subsidence would increase the temperature in the lower stratosphere; the reverse would apply for a high tropopause. Presumably the moderate correlation between $P_T$ and temperature in the lower stratosphere in winter and spring arises mainly from this effect of vertical motion. The surprising fact, however, is that although $P_T$ and $H_{300}$ are still moderately correlated in January and April the $H_{300}$ correlations with stratospheric temperatures are lower than the $P_T$ correlations. Table VII shows the correlations between $H_{300}$ and $P_T$ for the data used in preparing Table II and these may be compared with similar coefficients for larger numbers of observations in M.R.P. 819, Table II; it is seen that the correlations are always about -0.7 to -0.8. There is nothing inconsistent in the statistics, however, since a correlation coefficient of 0.7 merely signifies roughly a 50% relation, assuming the relationship to be linear. All the factors resulting in the reduced $H_{300}$ correlations in winter and spring at Lerwick are not obvious from this analysis, and further work is required to throw light on this interesting point.

In M.R.P. 819 the seasonal variation of ozone was compared with the seasonal variation of these correlation coefficients at Lerwick and Larkhill. The more reliable figures given here in Tables I, II and V certainly show that the correlation coefficients for the level 60 mb. (19 km.) which is not far below the ozone levels (20-35 km.) are, in general, least in January and April and greatest in October. The ozone maximum is in March and the minimum in October (see, for example, reference (3)) and the seasonal variation in the above correlation coefficients may thus well be associated with the variation in ozone, but this is not proved and the mechanism relating to it is obscure.

/Acknowledgment

## Acknowledgment

The correlation coefficients used in this analysis were computed by Mr. J.W. Bird.

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## TABLE I

### Stratospheric Statistics for Larkhill (51° 12'N, 01° 48' W)

### Period 1944-1952 (-1953 for January)

| | $H_{300}$ and T | | | $P_T$ and T | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $\bar{H}_{300}$ (10's) of ft.) | $\sigma H_{300}$ (10's of ft.) | $\rho$ | $\bar{P}_T$ (mb.) | $\sigma P_T$ (mb.) | $\bar{T}$ (°F.-ve) | $\sigma_T$ (°F.) | n |
| | | | | JANUARY | | | | | |
| Night 00,03 GMT | | | | | | | | | |
| $T_{150}$ | -0.61 | 2920 | 63 | 0.67 | 220 | 43 | 76.3 | 10.8 | 52 |
| $T_{100}$ | -0.53 | 2920 | 63 | 0.50 | 220 | 43 | 78.5 | 6.9 | 52 |
| $T_{60}$ | -0.28 | 2920 | 63 | 0.29 | 220 | 43 | 80.4 | 7.1 | 52 |
| Day 12,15 GMT | | | | | | | | | |
| $T_{150}$ | -0.60 | 3000 | 60 | 0.69 | 212 | 43 | 73.6 | 10.1 | 62 |
| $T_{100}$ | -0.67 | 3000 | 60 | 0.63 | 212 | 43 | 73.4 | 8.4 | 62 |
| $T_{60}$ | -0.27 | 3000 | 60 | 0.35 | 212 | 43 | 74.8 | 8.7 | 62 |
| | | | | APRIL | | | | | |
| Night 00,03,21 GMT | | | | | | | | | |
| $T_{150}$ | -0.77 | 3000 | 58 | 0.80 | 233 | 43 | 67.6 | 7.2 | 55 |
| $T_{100}$ | -0.67 | 3000 | 58 | 0.71 | 233 | 43 | 69.4 | 5.7 | 55 |
| $T_{60}$ | -0.22 * | 3000 | 58 | 0.37 | 233 | 43 | 69.9 | 5.0 | 55 |
| Day 12,15 GMT | | | | | | | | | |
| $T_{150}$ | -0.69 | 3010 | 58 | 0.74 | 230 | 45 | 63.4 | 7.2 | 89 |
| $T_{100}$ | -0.64 | 3010 | 58 | 0.61 | 230 | 45 | 63.8 | 5.7 | 89 |
| $T_{60}$ | -0.24 | 3010 | 58 | 0.39 | 230 | 45 | 61.5 | 5.7 | 89 |

| | $H_{300}$ and T | | | $P_T$ and T | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $\bar{H}_{300}$ (10's of ft.) | $\sigma H_{300}$ (10's of ft.) | $\rho$ | $\bar{P}_T$ (mb.) | $\sigma P_T$ (mb.) | $\bar{T}$ (°F.-ve) | $\sigma_T$ (°F.) | n |
| | | | | JULY | | | | | |
| Night 00,03 GMT | | | | | | | | | |
| $T_{150}$ | -0.86 | 3090 | 45 | 0.82 | 221 | 45 | 63.4 | 7.7 | 41 |
| $T_{100}$ | -0.79 | 3090 | 45 | 0.65 | 221 | 45 | 64.3 | 4.9 | 41 |
| $T_{60}$ | -0.55 | 3090 | 45 | 0.41 | 221 | 45 | 61.7 | 4.0 | 41 |
| Day 12,15 GMT | | | | | | | | | |
| $T_{150}$ | -0.71 | 3100 | 38 | 0.72 | 212 | 29 | 58.6 | 7.6 | 73 |
| $T_{100}$ | -0.70 | 3100 | 38 | 0.61 | 212 | 29 | 58.1 | 6.4 | 73 |
| $T_{60}$ | -0.40 | 3100 | 38 | 0.44 | 212 | 29 | 51.4 | 6.5 | 73 |
| | | | | OCTOBER | | | | | |
| Night 00,03 GMT | | | | | | | | | |
| $T_{150}$ | -0.80 | 3030 | 70 | 0.72 | 222 | 57 | 72.4 | 9.2 | 56 |
| $T_{100}$ | -0.73 | 3030 | 70 | 0.61 | 222 | 57 | 74.5 | 7.6 | 56 |
| $T_{60}$ | -0.56 | 3030 | 70 | 0.40 | 222 | 57 | 72.8 | 5.2 | 56 |
| Day 12,15 GMT | | | | | | | | | |
| $T_{150}$ | -0.88 | 3040 | 62 | 0.82 | 219 | 46 | 69.0 | 8.4 | 79 |
| $T_{100}$ | -0.79 | 3040 | 62 | 0.69 | 219 | 46 | 69.8 | 7.0 | 79 |
| $T_{60}$ | -0.57 | 3040 | 62 | 0.49 | 219 | 46 | 66.7 | 5.8 | 79 |

$H_{300}$ = height of 300 mb. surface

$P_T$ = pressure at the tropopause

$\sigma H_{300}$ = standard deviation of $H_{300}$

$\sigma_T$ = standard deviation of T

n = No. of observations

$T_{150}$, $T_{100}$, $T_{60}$ = temperature (°F.) at 150, 100, 60 mb. respectively

$\rho$ = appropriate correlation coefficient

$\sigma_{P_T}$ = standard deviation of $P_T$.

* Not significant at 5% level

## TABLE II

### Stratospheric Statistics for Lerwick ($60^\circ$ 08'N, $01^\circ$ 11' W)

Period 1944-1952 (-1953 for January)

|  | $H_{300}$ and T |  |  | $P_T$ and T |  |  |  |  |  |  | $H_{300}$ and T |  |  | $P_T$ and T |  |  |  |  |  |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
|  | $\rho$ | $\bar{H}_{300}$ (10's) of ft.) | $\sigma_{H_{300}}$ (10's of ft.) | $\rho$ | $\bar{\bar{P}}_T$ (mb.) | $\sigma_{P_T}$ (mb.) | $\bar{T}$ ($^\circ$F.-ve) | $\sigma_T$ ($^\circ$F.) | n |  | $\rho$ | $\bar{H}_{300}$ (10's of ft.) | $\sigma_{H_{300}}$ (10's of ft.) | $\rho$ | $\bar{P}_T$ (mb.) | $\sigma_{P_T}$ (mb.) | $\bar{T}$ ($^\circ$F.-ve) | $\sigma_T$ ($^\circ$F.) | n |
| Night<br>00,03,06,18 and 21 GMT | JANUARY |  |  |  |  |  |  |  |  | Night<br>00 GMT | JULY |  |  |  |  |  |  |  |  |
| $T_{150}$ | -0.58 | 2910 | 71 | 0.74 | 242 | 55 | 72.8 | 9.3 | 88 | $T_{150}$ | -0.26* | 3040 | 32 | 0.49 | 224 | 21 | 57.3 | 4.9 | 22 |
| $T_{100}$ | -0.30 | 2910 | 71 | 0.53 | 242 | 55 | 75.9 | 8.3 | 88 | $T_{100}$ | -0.32* | 3040 | 32 | 0.33* | 224 | 21 | 57.9 | 3.6 | 22 |
| $T_{60}$ | -0.11* | 2910 | 71 | 0.32 | 242 | 55 | 81.2 | 9.1 | 88 | $T_{60}$ | +0.07* | 3040 | 32 | -0.11* | 224 | 21 | 55.1 | 3.2 | 22 |
| Day<br>12,15 GMT |  |  |  |  |  |  |  |  |  | Day<br>12,15 GMT |  |  |  |  |  |  |  |  |  |
| $T_{150}$ | -0.19* | 2930 | 71 | 0.46 | 230 | 39 | 72.4 | 8.8 | 46 | $T_{150}$ | -0.79 | 3050 | 55 | 0.77 | 242 | 38 | 50.8 | 7.6 | 70 |
| $T_{100}$ | +0.08* | 2930 | 71 | 0.19* | 230 | 39 | 74.7 | 7.9 | 46 | $T_{100}$ | -0.71 | 3050 | 55 | 0.61 | 242 | 38 | 49.0 | 5.4 | 70 |
| $T_{60}$ | +0.22* | 2930 | 71 | 0.03* | 230 | 39 | 78.4 | 10.2 | 46 | $T_{60}$ | -0.39 | 3050 | 55 | 0.27 | 242 | 38 | 44.2 | 4.4 | 70 |
| Night<br>00,03,21 GMT | APRIL |  |  |  |  |  |  |  |  | Night<br>00,03 GMT | OCTOBER |  |  |  |  |  |  |  |  |
| $T_{150}$ | -0.70 | 2950 | 75 | 0.79 | 260 | 53 | 59.8 | 7.3 | 19 | $T_{150}$ | -0.81 | 3010 | 68 | 0.72 | 217 | 47 | 71.2 | 9.5 | 61 |
| $T_{100}$ | -0.73 | 2950 | 75 | 0.73 | 260 | 53 | 62.1 | 5.1 | 19 | $T_{100}$ | -0.76 | 3010 | 68 | 0.64 | 217 | 47 | 71.1 | 7.9 | 61 |
| $T_{60}$ | -0.40* | 2950 | 75 | 0.49 | 260 | 53 | 63.9 | 4.6 | 19 | $T_{60}$ | -0.57 | 3010 | 68 | 0.50 | 217 | 47 | 71.5 | 5.8 | 61 |
| Day<br>12,15 GMT |  |  |  |  |  |  |  |  |  | Day<br>12,15 GMT |  |  |  |  |  |  |  |  |  |
| $T_{150}$ | -0.54 | 2940 | 66 | 0.58 | 277 | 55 | 55.7 | 7.8 | 75 | $T_{150}$ | -0.83 | 3000 | 75 | 0.78 | 232 | 53 | 65.8 | 10.0 | 91 |
| $T_{100}$ | -0.35 | 2940 | 66 | 0.54 | 277 | 55 | 57.7 | 6.3 | 75 | $T_{100}$ | -0.82 | 3000 | 75 | 0.76 | 232 | 53 | 65.6 | 7.7 | 91 |
| $T_{60}$ | -0.07* | 2940 | 66 | 0.28 | 277 | 55 | 57.3 | 6.0 | 75 | $T_{60}$ | -0.61 | 3000 | 75 | 0.55 | 232 | 53 | 64.5 | 6.1 | 91 |

For explanation of symbols, see Table I

* Not significant at 5% level

TABLE III - Comparison of sets of upper air data: Larkhill

| | $\bar{H}_{300}$ (10's of ft.) | $\sigma H_{300}$ (10's of ft.) | $\bar{P}_T$ (mb.) | $\sigma P_T$ (mb.) | $\bar{T}_{150}$ ($-^{\circ}F.$) | $\sigma T_{150}$ ($^{\circ}F.$) | n |
|---|---|---|---|---|---|---|---|
| January | | | | | | | |
| Selected night ascents | 2920 | 63 | 220 | 43 | 76.3 | 10.8 | 52 |
| Selected day ascents | 3000 | 60 | 212 | 43 | 73.6 | 10.1 | 62 |
| All observations 1946-1950 * | 2969 | 77 | 236 | 54 | 68.9 | 9.7 | 528 ⌿ |
| April | | | | | | | |
| Selected night ascents | 3000 | 58 | 233 | 43 | 67.6 | 7.2 | 55 |
| Selected day ascents | 3010 | 58 | 230 | 45 | 63.4 | 7.2 | 89 |
| All observations 1946-1950 * | 2979 | 67 | 243 | 56 | 63.3 | 8.9 | 467⌿ |
| July | | | | | | | |
| Selected night ascents | 3090 | 45 | 221 | 45 | 63.4 | 7.7 | 41 |
| Selected day ascents | 3100 | 38 | 212 | 29 | 58.6 | 7.6 | 73 |
| All observations 1946-1950 * | 3081 | 44 | 223 | 39 | 60.1 | 7.7 | 560 ⌿ |
| October | | | | | | | |
| Selected night ascents | 3030 | 70 | 222 | 57 | 72.4 | 9.2 | 56 |
| Selected day ascents | 3040 | 62 | 219 | 46 | 69.0 | 8.4 | 79 |
| All observations 1946-1950 * | 3041 | 66 | 219 | 50 | 73.0 | 8.8 | 574 ⌿ |

For explanation of symbols, see Table I.

⌿ Number of observations of $T_{150}$.

* From M.O. 555, Part 1; statistics for $H_{300}$ refer to 1948-1950 only.

TABLE IV - Comparison of sets of upper air data: Lerwick

| | $\bar{H}_{300}$ (10's of ft.) | $\sigma H_{300}$ (10's of ft.) | $\bar{P}_T$ (mb.) | $\sigma P_T$ (mb.) | $\bar{T}_{150}$ ($-^{\circ}F.$) | $\sigma T_{150}$ ($^{\circ}F.$) | n |
|---|---|---|---|---|---|---|---|
| January | | | | | | | |
| Selected night ascents | 2910 | 71 | 242 | 55 | 72.8 | 9.3 | 88 |
| Selected day ascents | 2930 | 71 | 230 | 39 | 72.4 | 8.8 | 46 |
| All observations 1946-1950 * | 2895 | 76 | 254 | 57 | 69.0 | 9.5 | 502 ⌿ |
| April | | | | | | | |
| Selected night ascents | 2950 | 75 | 260 | 53 | 59.8 | 7.3 | 19 |
| Selected day ascents | 2940 | 66 | 277 | 55 | 55.7 | 7.8 | 75 |
| All observations 1946-1950 * | 2912 | 65 | 281 | 57 | 55.2 | 9.1 | 489 ⌿ |
| July | | | | | | | |
| Selected night ascents | 3040 | 32 | 224 | 21 | 57.3 | 4.9 | 22 |
| Selected day ascents | 3050 | 55 | 242 | 38 | 50.8 | 7.6 | 70 |
| All observations 1946-1950 * | 3033 | 51 | 247 | 37 | 51.0 | 7.0 | 541 ⌿ |
| October | | | | | | | |
| Selected night ascents | 3010 | 68 | 217 | 47 | 71.2 | 9.5 | 61 |
| Selected day ascents | 3000 | 75 | 232 | 53 | 65.8 | 10.0 | 91 |
| All observations 1946-1950 * | 2967 | 71 | 234 | 53 | 66.9 | 9.9 | 549 ⌿ |

For explanation of symbols, see Table I.

⌿ Number of observations of $T_{150}$.

* From M.O. 555, Part 2; statistics for $H_{300}$ refer to 1948 -1950 only.

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## TABLE V - Stratospheric statistics for Lerwick in January
## Day and night combined: 1944-53

| | $H_{300}$ and T | | | $P_T$ and T | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $H_{300}$ (10's of ft.) | $\sigma H_{300}$ | $\rho$ | $\bar{P}_T$ (mb.) | $\sigma P_T$ (mb.) | $\bar{T}$ (-°F.) | $\sigma_T$ (°F.) | n |
| $T_{150}$ | -0.45 | 2920 | 71 | 0.66 | 238 | 50 | 72.6 | 9.1 | 134 |
| $T_{100}$ | $-0.16^*$ | 2920 | 71 | 0.43 | 238 | 50 | 75.5 | 8.1 | 134 |
| $T_{60}$ | $+0.03^*$ | 2920 | 71 | 0.21 | 238 | 50 | 80.2 | 9.9 | 134 |

For explanation of symbols, see Table I.

* Not significant at 5% level.

## TABLE VI - Ratio of squares of correlation coefficients for $T_{60}$ and $T_{150}$

| | January | | April | | July | | October | | Year // | |
|---|---|---|---|---|---|---|---|---|---|---|
| | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ |
| **Larkhill** | | | | | | | | | | |
| Night | 0.21 | 0.19 | $0.08^*$ | 0.21 | 0.41 | 0.25 | 0.49 | 0.31 | | |
| Day | 0.20 | 0.26 | 0.12 | 0.28 | 0.32 | 0.37 | 0.42 | 0.36 | 0.25 | 0.31 |
| **Lerwick** | | | | | | | | | | |
| Night | $0.03^*$ | 0.19 | $0.33^*$ | 0.38 | $0^*$ | $0.05^*$ | 0.49 | 0.48 | | |
| Day | $0^*$ | $0.004^*$ | $0.02^*$ | 0.23 | 0.24 | 0.12 | 0.54 | 0.50 | 0.12 | 0.21 |

* Correlation for $T_{60}$ (or $T_{150}$) not significant at 5% level. If correlation coefficient changes sign from 150 to 60 mb., ratio counted as zero.

// Values here are the squares of the mean ratios of the daylight coefficients for $T_{60}$ and $T_{150}$ for all four months, except for January, Lerwick where the correlation coefficients for all hours combined were used instead of the daylight values.

## TABLE VII - Correlation coefficients between $H_{300}$ and $P_T$
## High altitude ascents at Lerwick

| | Night | | Day | |
|---|---|---|---|---|
| | Correlation Coefficient | No. of Obs. | Correlation Coefficient | No. of Obs. |
| January | -0.77 | 88 | -0.87 | 46 |
| April | -0.85 | 19 | -0.79 | 75 |
| July | -0.69 | 22 | -0.86 | 69 |
| October | -0.75 | 61 | -0.83 | 91 |

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M.R.P. 820

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# AIR MINISTRY

# METEOROLOGICAL RESEARCH COMMITTEE

## Five Flights through a Thunderstorm Belt
by
[illegible] (Meteorological Office)

### 1. Introduction

In a previous report (1) on accelerometer investigation the writer gave the results of some integrations of the accelerometer records obtained in a Spitfire aircraft on flights through cumulus and cumulonimbus clouds. The curves obtained, curves of vertical speed, required adjustment to take account of any errors in reading the accelerometer trace and it was necessary to make certain plausible assumptions, as to the changes of height of the aircraft, to obtain reasonable curves of the vertical speeds to which the aircraft was subjected. Lack of information of height changes experienced precluded any check on the validity of these assumptions.

With the change of aircraft type for these investigations from Spitfire to [illegible] it became possible to carry an observer and to install an auto-recorder to give photographically at 10 second intervals the readings of I.C.A.N. height, air temperature and indicated air speed. Any accelerometer record from the [illegible], therefore, can be integrated and the curve obtained adjusted to ensure that the vertical speeds obtained give, by a further integration, correct height changes over successive periods of 10 seconds or more.

By integrating further accelerometer records obtained in the [illegible] it is therefore possible to check whether the assumptions made in the previous report were reasonable as well as obtain information on vertical currents which will be of interest on their own account.

Five flights made on 12th June, 1952 through a belt of thunderstorms were chosen for analysis in this way and are the subject of this report.

### 2. Synoptic situation and flight programme

The [illegible] of England and Wales was covered by a [illegible]

[illegible]

[illegible]

| | $H_{300}$ and T | | | | $P_T$ and T | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | $\rho$ | $H_{300}$ (10's of ft.) | $\sigma H_{300}$ | | $\rho$ | $\bar{P}_T$ (mb.) | $\sigma P_T$ (mb.) | $\bar{T}$ (-°F.) | $\sigma T$ (°F.) | n |
| $T_{150}$ | -0.45 | 2920 | 71 | | 0.66 | 238 | 50 | 72.6 | 9.1 | 134 |
| $T_{100}$ | -0.16* | 2920 | 71 | | 0.43 | 238 | 50 | 75.5 | 8.1 | 134 |
| $T_{60}$ | +0.03* | 2920 | 71 | | 0.21 | 238 | 50 | 80.2 | 9.9 | 134 |

For explanation of symbols, see Table I.

* Not significant at 5% level.

## TABLE VI - Ratio of squares of correlation coefficients for $T_{60}$ and $T_{150}$

| | January | | April | | July | | October | | Year // | |
|---|---|---|---|---|---|---|---|---|---|---|
| | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ | $H_{300}$ | $P_T$ |
| **Larkhill** | | | | | | | | | | |
| Night | 0.21 | 0.19 | 0.08* | 0.21 | 0.41 | 0.25 | 0.49 | 0.31 | | |
| Day | 0.20 | 0.26 | 0.12 | 0.28 | 0.32 | 0.37 | 0.42 | 0.36 | 0.25 | 0.31 |
| **Lerwick** | | | | | | | | | | |
| Night | 0.03* | 0.19 | 0.33* | 0.38 | 0* | 0.05* | 0.49 | 0.48 | | |
| Day | 0* | 0.004* | 0.02* | 0.23 | 0.24 | 0.12 | 0.54 | 0.50 | 0.12 | 0.21 |

* Correlation for $T_{60}$ (or $T_{150}$) not significant at 5% level. If correlation coefficient changes sign from 150 to 60 mb., ratio counted as zero.

// Values here are the squares of the mean ratios of the daylight coefficients for $T_{60}$ and $T_{150}$ for all four months, except for January, Lerwick where the correlation coefficients for all hours combined were used instead of the daylight values.

## TABLE VII - Correlation coefficients between $H_{300}$ and $P_T$
## High altitude ascents at Lerwick

| | Night | | Day | |
|---|---|---|---|---|
| | Correlation | No. of | Correlation | No. of |